PAN-AFRICAN IMPRINT ON THE EARLY TO MID-PROTEROZOIC RICHTERSVELD AND BUSHMANLAND SUBPROVINCES NEAR EKSTEENFONTEIN, NAMAQUALAND, REPUBLIC OF SOUTH AFRICA

by

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Thesis submitted in fulfilment of the requirements for the degree of Doctor of Philosophy in the Faculty of Science of the University of Cape Town

Department of Geology

1990

ABSTRACT

The present investigation examines the relationship between the Proterozoic Richtersveld and Bushmanland Subprovinces in the westernmost part of the Namaqua Province, near Eksteenfontein, Republic of South Africa. There is a controversy about this relationship because isotopic data contrast with field evidence. On a regional scale the Richtersveld Subprovince is separated from the Bushmanland Subprovince by the northward-dipping Groothoek Thrust. North of the thrust the Richtersveld Subprovince is comprised of low grade volcano/ plutonic rocks of the Vioolsdrif Terrane and medium grade volcano sedimentary sequences of the Pella Terrane. Medium grade rocks of the Steinkopf Terrane (Bushmanland Subprovince) lie immediately south of the thrust. Late Proterozoic strata of the Stinkfontein Formation (Gariep Group) overlie the Namaqua Province in the west; Cambrian Nama Group outliers occur east of the Stinkfontein Formation.

Isotopic data show that lithologies of the Richtersveld Subprovince formed between 2000 - 1730 Ma, whereas those of the Bushmanland Subprovince are younger. It is not clear whether the Namaqua metamorphic imprint (at 1200 - 1100 Ma), which is manifest in terranes south of the Groothoek Thrust, extended as far as the Vioolsdrif Terrane in the north. Early Proterozoic structural and metamorphic imprints are inferred to have been obliterated during this event. The westernmost part of the Namaqua Province was overprinted for a distance of 100 km from the coast, during the Pan-African event at 700 Ma and 500 Ma.

An area measuring nearly 500 km^2, traversing the western extremity of the boundary between the Richtersveld and Bushmanland Subprovinces was mapped on a scale of 1:36,000. Field mapping was carried out with the aid of aerial photographs, whereas laboratory techniques included map compilation, structural analysis, X-ray diffractometry, geochemical (XRF) and electron microprobe analyses.

Supracrustal units of the Richtersveld Subprovince are composed of quartzo-feldspathic gneisses, schists, and minor meta-pelites. Supracrustals of the Bushmanland Subprovince are less diverse than those of the Richtersveld Subprovince and have a disconformable relationship with them. Most intrusive rock-types are thick granitic sheets, except the Early Proterozoic Vioolsdrif Granodiorite which forms part of a batholithic pluton in the north. The Sabieboomrante adamellite gneiss, Kouefontein granite gneiss and Dabbieputs granite gneiss could not be correlated with lithologies commonly occurring in the Richtersveld and Bushmanland Subprovinces. They have been given the new rock names. Mafic and ultramafic rocks of the Klipbok complex occur along the strike of the Groothoek Thrust. They form part of the Richtersveld Subprovince.

The west-northwest Namaqua structural trend is repeatedly offset by Late Proterozoic/Early Palaeozoic (Pan-African) mega-shears almost orthogonal to it. The sense of shear along these zones is predominantly left-lateral, strike-slip, but also upthrown to the west. Most shears have, in addition, a rotational component. Rock units are disrupted along strike resulting in an irregular outcrop pattern. Eight thrust zones, of which the Ratelfontein and Groothoek Thrust zones are the most important, occur parallel to the regional Namaqua structural grain. They are mostly of Early to Mid Proterozoic age. Strata are imbricated in the Ratelfontein Thrust sheet; re-orientation of folds and lineations, shearing, and mylonite development over a width of 2 km are associated with the Groothoek Thrust.

In the eastern part of the study area there is an increase in metamorphic grade from upper greenschist facies in the north to upper amphibolite facies in the south-east. This pattern relates to the first Namaqua metamorphic episode; temperatures reached 778 ± 50°C, whereas pressures are estimated at 3.5 kbar. The second Namaqua metamorphic episode took place under greenschist facies conditions. The first Pan-African metamorphic episode is characterized by staurolite-kyanite assemblages in the western part of the study area, and greenschist facies assemblages in the south-east. Temperatures during this metamorphic episode reached 630 ± 50°C, and pressures were 5.5 kbar. Subsequent metamorphic imprints have muscovite-chlorite-sericite assemblages and are confined to Pan-African shear zones.

Tectonic models incorporating subduction and arc accretion are proposed to account for Early Proterozoic (pre-Namaqua) deformation and metamorphism. Arc accretion is thought to have occurred at about 1850 Ma, the Ratelfontein Thrust forming a cryptic suture. Mafic and ultramafic rocks of the Klipbok complex may represent dismembered ophiolites, although field and geochemical evidence do not conclusively support this interpretation. A model of southwestward movement along thrusts at deep crustal levels, accompanied by emplacement of granitic sheets, is proposed for the Mid-Proterozoic history of the north-west Namaqua Province. Groothoek thrusting took place at mid-crustal levels just prior to 1100 Ma. Metamorphism outlasted deformation during the latter episode, taking place under conditions of lowering pressure. It completely obliterated the earlier metamorphic history.

A model incorporating strike-slip movement along shears in the basement contemporaneous with thrusting in cover rocks is proposed for the first Pan-African episode. In this model tectonic loading induced by south-easterly transported nappes caused metamorphism to lower amphibolite facies and medium pressure conditions in the western part of the area. Deformation and metamorphism thereafter were at shallower crustal levels; this was concentrated along steeply-dipping shear zones, the westernmost shears transecting strata of the Stinkfontein Formation. Subsequent movement mainly along the Steenbok Shear deformed sediments of the Nama Group and developed brittle zone structures, under greenschist facies conditions.

The Eksteenfontein area is interpreted as forming a recess, whilst low-angle thrusting took place within Gariep cover rocks to the north-west, and later in Nama cover rocks to the south-east of the area.

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PLATES

Mineral abbreviations used in this text:

ab	= albite	ga	= garnet
act	= actinolite	hb	= hornblende
an	= anorthite	Kf	= K-feldspar
and	= andalusite	ky	= kyanite
ant	= anthophyllite	mu	= muscovite
br	= brucite	micr	= microcline
bt	= biotite	phl	= phlogopite
cc	= calcite	plag	= plagioclase
cd	= corundum	qtz	= quartz
chd	= chloritoid	ser	= sericite
chl	= chlorite	sil	= sillimanite
crd	= cordierite	sp	= serpentine
czo	= clinozoisite	st	= staurolite
en	= enstatite	stp	= stilpnomelane
ep	= epidote	tr	= tremolite
fo	= forsterite		

1 INTRODUCTION

1.1 AIMS OF PRESENT INVESTIGATION

The investigation examines (i) the relationship between the Early to Mid- Proterozoic Richtersveld and Bushmanland Subprovinces of the Namaqua Province at the western extremity of the zone separating the two subprovinces (Fig. 1), and (ii) the effects of Late Proterozoic - Early Palaeozoic (Pan-African) tectonism and metamorphism on these rocks.

The relationship between the Richtersveld and Bushmanland Subprovinces in the extreme North Western Cape is a topic of conjecture. Some investigators (Ward, 1977; Ritter, 1980) interpret the boundary between the two subprovinces as conformable. In these models rocks of the Richtersveld Subprovince are presumed younger than those of the Bushmanland Subprovince. Other investigators propose the opposite situation, based on field and radiometric evidence (Theart, 1980; Strydom, 1982, 1985; Blignault et al., 1983; Reid and Barton, 1983; Hartnady et al., 1985). The latter interpretation implies an overthrust relationship between the two subprovinces in which lithologies of the Richtersveld Subprovince are juxtaposed across those of the Bushmanland Subprovince (Blignault et al., 1983; Van der Merwe, 1986; Van Aswegen, 1988).

The study area is situated immediately east of the Pan-African Gariep Belt and Late Proterozoic - Early Palaeozoic (Pan-African) tectonism and metamorphism are overprinted along a 30 km wide marginal zone (De Villiers and Söhnge, 1959; Joubert, 1971; Ritter, 1980; Waters, Joubert and Moore, 1983). This factor complicates interpretations of the earlier geologic history.

The aims of the present investigation therefore are to:

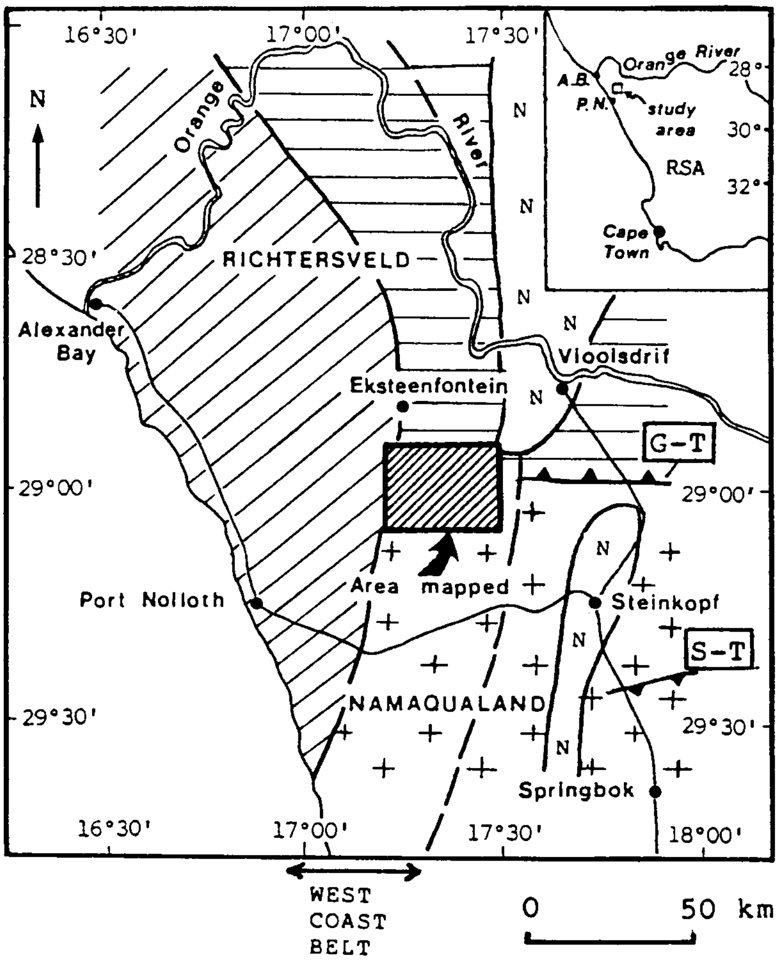
- Identify rock-types belonging to the Bushmanland and Richtersveld Subprovinces in their westernmost outcrop area;

- Determine whether the nature of the boundary between these two subprovinces is a normal stratigraphic, or tectonic one;

- Carry out a structural analysis to establish a sequence of Early to Mid-Proterozoic and Late Proterozoic to Early Palaeozoic (Pan-African) deformational events;

- Study key metamorphic rocks to determine the grade of metamorphism associated with each of the deformational events;

- Propose geotectonic models to account for the Early to Mid-Proterozoic and Late Proterozoic (Pan-African) tectonic and metamorphic history of the region.



N Nama Group (Cambrian)

Gariep Belt (Late Proterozoic)

Richtersveld Subprovince (Early Proterozoic) } Namaqua Province

Bushmanland Subprovince (Early? to Mid-Proterozoic) } Namaqua Province

Major thrust zone; G-T = Groothoek Thrust

S-T = Skelmfontein Thrust

Fig. 1 Locality Map of area investigated.

1.2 PREVIOUS AND OTHER RELEVANT WORK

Previous workers (e.g. De Villiers and Söhnge, 1959; Joubert, 1971; Ritter, 1980) whose surveys covered the present area of interest terminated their investigations along the 29th parallel. This was largely due to the nature of their regional geological reconnaissance-type mapping surveys where boundaries were defined quite often by lines of latitude and longitude. North of the 29th parallel De Villiers and Söhnge (1959) mapped the eastern Richtersveld up to the Orange River and produced the first base map of this region.

South of the 29th parallel Joubert (1971) mapped a large area (2000 km^2) down to 30°30′, providing a foundation for the understanding of rock types, metamorphism and structural evolution of this complex area.

Ritter (1980) remapped the entire eastern Richtersveld in order to gain a better understanding of lithostratigraphy, structure and metamorphism of this region - especially with regard to its relationship with gneissic rocks occurring to the immediate south, in Namaqualand.

Other relevant work, but not in the immediate area of investigation, includes regional surveys to the east by Ward (1977), Theart (1980), Van Aswegen (1981, 1988), Strydom (1985) and Van der Merwe (1986). West of the study area, Kröner (1974) investigated rock types of the Gariep Group and their tectogenesis. Middlemost (1963) carried out a detailed examination of predominantly igneous rocks in the south-eastern Richtersveld, just to the north of the present area of investigation.

Reconnaissance radiometric dating of lithologies in an extensive area of the Namaqua mobile belt was carried out by Nicolaysen and Burger (1965). More detailed geochronologic studies specific to the area under consideration were carried out by De Villiers and Burger (1967), Allsopp et al. (1979), Welke et al. (1979) and Reid (1979a and b; 1981).

1.3 STUDY AREA, LOCALITY, ACCESS AND PHYSIOGRAPHY

The study area lies approximately 30 km south-east of Eksteenfontein, on the southern fringes of a geographic region known as the Richtersveld (Fig. 1). During my study boundaries were arbitrarily chosen to include rock-types typical of the Richtersveld and Bushmanland Subprovinces. Field mapping was carried out using aerial photographs on a scale of 1:36,000. The study area covers three flight lines, viz. strips 16, 17 and 18 of Job Number 525. The area mapped measures nearly 500 km^2; the approximate west to east distance being 27 km, and north to south, 18 km.

The western boundary coincides with the Late Proterozoic Stinkfontein Formation (Gariep Group), whereas Cambrian Nama Group cover rocks (forming part of the Neint Nababeep Plateau) conveniently delimit the eastern boundary. Where Nama Group rocks are absent, as in the south-east, the 17°30′ longitude line continues as the arbitrarily chosen eastern boundary.

Access to the study area from the south is via two good gravel roads, leading off from the main tar road between Port Nolloth and Steinkopf. From the study area a gravel road leads northwards to Eksteenfontein, and another westwards to Lekkersing. Over the rest of the area access roads are either old disused farm tracks, or prospectors tracks, most of which have experienced wash-aways and are in a constant state of disrepair. Due to the hilly nature of the topography, most of the area was surveyed on foot.

There is a dominant north-westerly trending watershed which traverses the northern half of the study area. This watershed divides drainage northwards into the Orange River from a south and south-westward drainage system into the Atlantic Ocean. Several rock-types are exposed along the watershed, but for a large part granitic rocks (Kouefontein granite gneiss) crop out at an average elevation of 860 m above sea level in the east (maximum 943 m at beacon number 19 on Kouefontein), and 890 m in the north-west (maximum 950 m at beacon number 42 on Kromnek).

The slopes facing away from the watershed towards the south and south-west form much steeper slope angles than those facing towards the Richtersveld in the north. This is a function of two factors, viz., rock-type resistance to weathering, and an erosional plain to the south at least 200 m lower than one near Eksteenfontein in the north. Although weathering processes are predominantly mechanical, both the above factors contribute to a slope retreat advancing more rapidly from the south.

The maximum relief is of the order of 650 m, predominantly a rugged hilly type of topography. The vegetation is a sparse low bush with very rare thorn trees. Natural succulent growth is largely dependant on the influx of moisture from the coast, usually as early morning mist.

A large portion in the west and south-west of the area is covered by scree and red wind-blown sand. Many of the streams end up as part of an internal drainage system because this is a low rainfall area, (<25 mm p.a.) (see Kabies se Pan and Kompan, grid reference G11 and I14, Annexure 1). Outcrop in this vicinity is confined to inselberg-type topography, e.g. Kabies se Berg.

To the north of the watershed outcrop is good, and continuous over most of the area. Nama Group sediments form the high ground in the east and north-east, the highest point being Bluff beacon (number 44) at an elevation of 986 m.

In the west, the basal Stinkfontein Formation forms a rugged topographic plateau, rising from 500 m in the south to just over 700 m in the north. Farther west and northwards from the study area Stinkfontein quartzites form a prominent mountain range towards Eksteenfontein and beyond.

1.4 METHODS OF STUDY

These include field mapping, compilation of maps, analysis of structural data, geochemical analyses by XRF and electron microprobe, and analysis by X-ray diffraction. Details of methods are given in Appendix A-1.

1.5 ACKNOWLEDGEMENTS

Prof. P. Joubert initiated the project and acted as supervisor from it's commencement in July, 1983, until his retirement at the end of 1986. I wish to thank him for making me aware of certain complexities regarding rock-type relationships in the Richtersveld and Bushmanland Subprovinces.

Prof. C.J. Hartnady took over supervision until completion of the project. His critical reviews and discussions of the original manuscript led to improvements, and a better understanding and knowledge of the complex geology of the Namaqualand and Richtersveld region. I would like to thank him for advice, and understanding many of my personal difficulties throughout the study period.

A revised draft of the manuscript was critically reveiwed by Prof. M.J. de Wit. I wish to sincerely thank him for taking an interest in my project and contributing valuable advice. My geologic knowledge has benefitted considerably through his influence.

I wish to thank Dr. D.J. Waters, first for his patience in assiting me with certain petrological problems and secondly, for granting permission to attend his metamorphic petrology course for honours students. My understanding of metamorphism improved substantially through his lectures and subsequent discussions.

Dr. D.L. Reid very kindly spent a week in the field with me, during which time I was able to learn an immense amount from his personal knowledge of rock-types in the Richtersveld and Namaqualand region. I benefitted from his observations with respect to the nature of rock-types in the study area, and his probing questions. I would like to thank him for carrying out geochemical analyses of plutonic rocks included in Tables 3.21, 3.24, 3.26 and 3.27. I learned much from him during discussions of the results.

I had very stimulating discussions with, and advice from many of the participants of the 1984 and 1987 Geocongress excursions to the Richtersveld and Namaqualand. In general their suggestions led to further investigations on my part, and this in turn opened up new avenues in my understanding of the subject of geology. I wish to thank them all for the part they played in assisting me to improve my geological knowledge.

Dr. A.E. Shimron provided me with useful insight into solving certain petrographic and structural problems. I would like to thank him for taking an interest in my progress and for several stimulating discussions.

I wish to thank Dr. J.M. Moore for carrying out geochemical analyses of pelitic rocks included in Table 3.6. He provided me with information on metapelites from his unpublished Ph.D. thesis which I have used in this text for comparative purposes.

I would like to thank Dr. M.K. Watkeys for fruitful discussions concerning the Groothoek Thrust, and for providing me with information on the location of the thrust in the Pofadder area.

Dr. K. Kasch very kindly reviewed the chapter on metamorphism. I am grateful to him for finding time outside of his lecturing contract period in the Geology Department at U.C.T. to make several very useful suggestions to improve the chapter.

My colleagues at the University of Port Elizabeth, Prof. I.C. Rust, Dr. R.W. Shone and Mr. G.W. Berg, always showed an interest in my progress. Their contributions by way of advice and discussions are gratefully acknowledged. Mr. Berg is thanked for instruction in the use of the X-ray diffractometer, and Miss. J. Burkinshaw for her assistance in determining unknown minerals on the machine.

Prof. R.E. Jacob, of Rhodes University, gave me exceptionally useful advice on several occasions. I wish to thank him for taking an interest in my project, and unselfishly devoting his time during several discussions.

Prof H.V. Eales is thanked for granting permission to use the electron microprobe at Rhodes University. Messers. R. Skae and R. Rickard were very patient in their instruction in the use of the electron microprobe. I wish to thank them both for being such good teachers.

The geological staff of the Okiep Copper Company, especially Mr. J. Marais and Mr. B. Packam, are thanked for their assistance during my initial field spell in Namaqualand.

De Beers Consolidated Mines Ltd. are thanked for permission to traverse the farm Oograbies.

To my wife Lynne a very big thank you for her tremendous moral support, and bearing many of the family responsibilities during an exceptionally demanding but rewarding period of study for me. She also performed the arduous task of typing countless notes, several reports and the original manuscript, thereby saving me a vast amount of time and lessening our financial burden substantially. Our daughters Cheryl and Vonny are thanked for their assistance, especially colouring in the maps.

I would like to thank the Venerable and Mrs. A.J. Kriel for showing a constant parental interest in my progress, and for their assistance in some of the printing costs of this thesis.

I appreciated the hospitality shown to me by Mr. and Mrs. J. Brand and Mr. and Mrs. H. Mostert and their families who reside on the farm "Tussen In". They were my only neighbours in that frontier part of Namaqualand and I would like to thank them for readily accepting me into their company, although I was a total stranger to them.

To the ex-headmaster of the Eksteenfontein Junior School, Mr. P. Cloete and his family, I wish to express my sincere thanks for always taking an interest in my progress, and extending a hand of friendship in that remote part of the Republic.

I would like to thank the University of Port Elizabeth for some financial support which assisted me in carrying out the research work.

Mrs. R. Kovats and Mrs S. Sayers are thanked for final drafting of the 1:50,000 and 1:30,000 geological maps (Annexures 1 and 2).

2 A REVIEW OF NAMAQUALAND/RICHTERSVELD - RELATED LITERATURE AND EXTRACTION OF CONJECTURAL TOPICS

2.1 INTRODUCTION.

A literature search was undertaken of publications pertaining to the north-western part of the Namaqua Province to ascertain and outline topics of conjecture. One of the topics extracted from this survey focuses on controversial interpretations regarding the relationship between the Richtersveld and Bushmanland Subprovinces An area in Namaqualand was then chosen where research could be directed at finding a solution to this problem.

The reader unfamiliar with the area and geologic terminology pertaining to the Proterozoic history of the 2000 - 1000 Ma Namaqua Province will find an explanation of the following areas and terms useful.

The **Richtersveld** is the geographical region confined to the extreme N.W. Cape (Fig 1). It extends southwards from the Orange River for about 100 km to just south of Eksteenfontein, and has a width of about 80 km. **Namaqualand** fringes the Richtersveld, extending inland from the coast for a distance of some 100 km. The Orange River forms its northern boundary and the Vanrhynsdorp area (some 200 km south-southeast of Springbok) is regarded as its southern boundary. The region east of Namaqualand is known as **Bushmanland.**

Namaqualand Metamorphic Complex (Botha, 1983) and **Namaqua mobile belt** (Joubert, 1981) are synonymous terms denoting lithostratigraphy, tectonism and metamorphism associated with the 2000 – 1000 Ma Namaqua Province (Table 2.1). For consistency the term **Namaqua mobile belt** is adhered to throughout this thesis.

"Namaqua geotraverse" refers to an area of the Namaqua Province investigated by Blignault et al. (1983), extending from Springbok in the south to just north of Vioolsdrif on the Orange River (Fig. 2.3).

The **Gariep Belt** is confined to the outcrop area of the Late Proterozoic Gariep Group which forms cover rocks to the Namaqua Province in the west (Fig. 2.3).

The **West Coast Belt** refers to the tectonic and metamorphosed marginal zone to the Gariep Belt, extending for some 100 km inland from the coast (Fig. 2.3). Rocks of the Early to Mid-Proterozoic Namaqua Province were overprinted in this belt during the Pan-African (Late Proterozoic-Early Palaeozoic) event.

Table 2.1 Geological nomenclature applied to Early to Mid-Proterozoic tectonostratigraphic terranes and subprovinces in the western part of the Namaqua Province

Province	Subprovince	Terrane	Lithological Group/Sequence	Lithological Subgroup	Formation	Suite	Metamorphic Grade	Age in Ma	Orogeny/ Deformation event
Namaqua	Richtersveld	Vioolsdrif	Orange River Group	Haib	Nous Tsams Windvlakte		Upper green-schist	2000 (Reid & Barton '83)	Eburnian
						Vioolsdrif		1900-1730 (Reid'79b)	
		Pella	Aggeneys Sequence (=Bushmanland Sequence)		Pella Kabas Dabenoris		amphibolite	1900-1600? (Köppel '82 Reid '81)	Namaqua (Kibaran)
			Orange River Group		Guadom Hom		amphibolite		
	Bushmanland	Steinkopf		Een Riet	Groothoek [=Transitional Zone (Theart '80)]		amphibolite	1800?	Namaqua (& 1800 event?)
						Gladkop		1800 (Barton et al. '81)	1800 event?
		Okiep			Khurisberg		granulite	1800?	Namaqua
						Little Namaqualand		1200 (Clifford et al. '81)	
						Spektakel		1100 (Clifford et al. '74)	

2.2 A BROAD PERSPECTIVE OF THE NAMAQUA MOBILE BELT

Early to Mid- Proterozoic metasediments, metavolcanics and intrusive rocks of the Namaqua mobile belt crop out in an elongate westerly-trending wedge-shaped area in the extreme north-west Cape and Namibia. They extend eastwards below younger Phanerozoic cover, and re-appear on the east coast in Natal (Joubert, 1981; Matthews, 1972; 1981). The mobile belt was originally delineated as a very extensive area of 1000 Ma rocks (Nicolaysen and Burger, 1965), but it is now recognized that this date represents the last tectonothermal veil during which isotopic clocks were "last" reset (Joubert, 1981). In the north-west Cape and Namibia a number of isotopic age dates reveal that this portion of the earth's crust first formed between 2000-1000 Ma (De Villiers and Burger, 1967; Welke et al., 1979; Reid, 1979b, 1981). The oldest dates are recorded from the Richtersveld Subprovince - a small low-grade metamorphic enclave in the extreme north-west of the Republic of South Africa and southern Namibia (Fig. 2.3). Surrounding this enclave the Namaqua mobile belt in the northern Cape and Namibia shows polyphase deformation, with a high proportion of granitic rocks, and accompanying high-temperature low-pressure metamorphism.

Gravity data indicate that the northern and eastern boundaries of the mobile belt with the Kaapvaal craton are abrupt; this is inferred as a consequence of large-scale uplift of the mobile belt (De Beer and Meyer, 1983). This interpretation is consistent with structural and metamorphic studies (Van Bever Donker, 1980; Van Zyl, 1981). The southern boundary is generally taken as the Beattie magnetic anomaly and the associated Southern Cape Conductive Belt (SCCB). This anomaly runs parallel to and some 300 km inland from the southern coast, and is reflected beneath the Phanerozoic Karoo cover (Van Zyl, 1981; De Beer and Meyer, 1983). The SCCB is interpreted as a zone of serpentinized basic rocks representing oceanic crust obducted from the south (De Beer and Meyer, 1983).

Early to Mid-Proterozoic deformational phases are recognized over the entire mobile belt. At least four main phases, with associated metamorphism, occur in the Namaqua mobile belt in the Republic of South Africa, but more deformation phases are present in the Namibian portion of the mobile belt (Joubert, 1981, p. 690). Generally, the earlier two deformation and metamorphic events were the most intense; later events being characterised by brittle deformation structures such as steeply-dipping shear zones and less intense folding.

Cover formations include:

- the Gariep Group, a 900 Ma north-south trending eugeoclinal/ miogeoclinal sequence, confined to the west coast. The Stinkfontein Formation (base of Gariep Group) forms the western boundary to the study area (Annexure 1);

- the Nama Group, a 600 Ma sequence of mostly undeformed clastic and limestone deposits. These are the main cover rocks in the north-east of the study area (Annexure 1). Nama Group rocks occur as outliers, and down-faulted blocks within the Namaqua mobile belt as far south as Vanrhynsdorp;

- the Karoo Supergroup, a Phanerozoic sequence of mainly continental sediments and volcanics forming the southern cover to the Namaqua mobile belt. These lithologies do not extend as far north as the study area.

2.3 TECTONIC PROVINCES IN WESTERN NAMAQUALAND.

2.3.1 Introduction

Kröner and Blignault (1976, p. 232) define a tectonic PROVINCE as "A geographic region that is characterized by a combination of such parameters as lithology, structure, metamorphism, and predominant radiometric age differing significantly from those of adjacent areas." In one of the earliest attempts at synthesizing available stratigraphic, structural, metamorphic, and radiometric data, these authors outline tectonic and igneous provinces in the North-west Cape/southern Namibia, based on criteria such as regional structural/metamorphic patterns, duration of tectonism, recognition of lithologies as supracrustal/platform sequences, and establishing levels of emplacement. The boundaries of tectonic provinces are demarcated by marginal zones. Thus the main tectonic provinces are defined by Kröner and Blignault (1976) as Fig. 2.1:

- the Gariep Province (700 - 900 Ma) (= outcrop area of Gariep Group);
- the Namaqua Province (>1200 Ma);
- the Richtersveld Province (1900 Ma);
- the Kheis Province;
- the Kaapvaal Province (≥2500 Ma).

A structural analysis of a portion of Northern Namaqualand carried out by Theart (1980) reveals significant discordant boundaries between structural domains. Some of these boundaries have subsequently been interpreted as tectonic boundaries (Blignault et al., 1983; Van der Merwe, 1986).

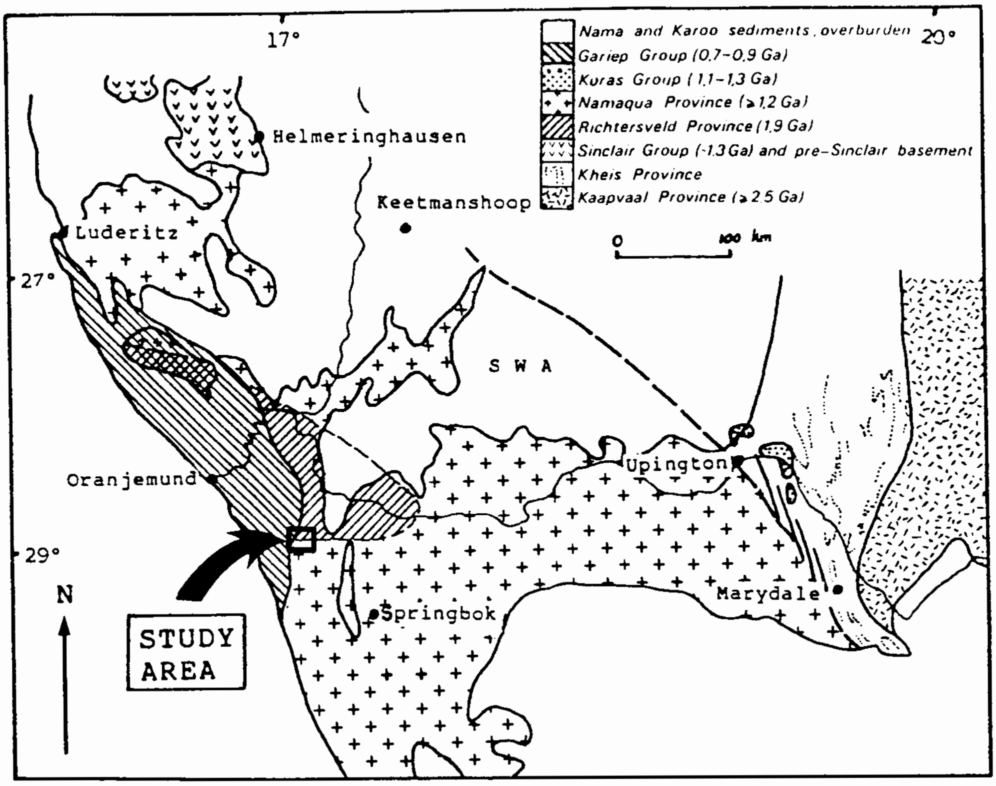


Fig. 2.1 Outline of the Richtersveld, Namaqua and Gariep tectonic provinces (after Kröner and Blignault, 1976, Fig. 1).

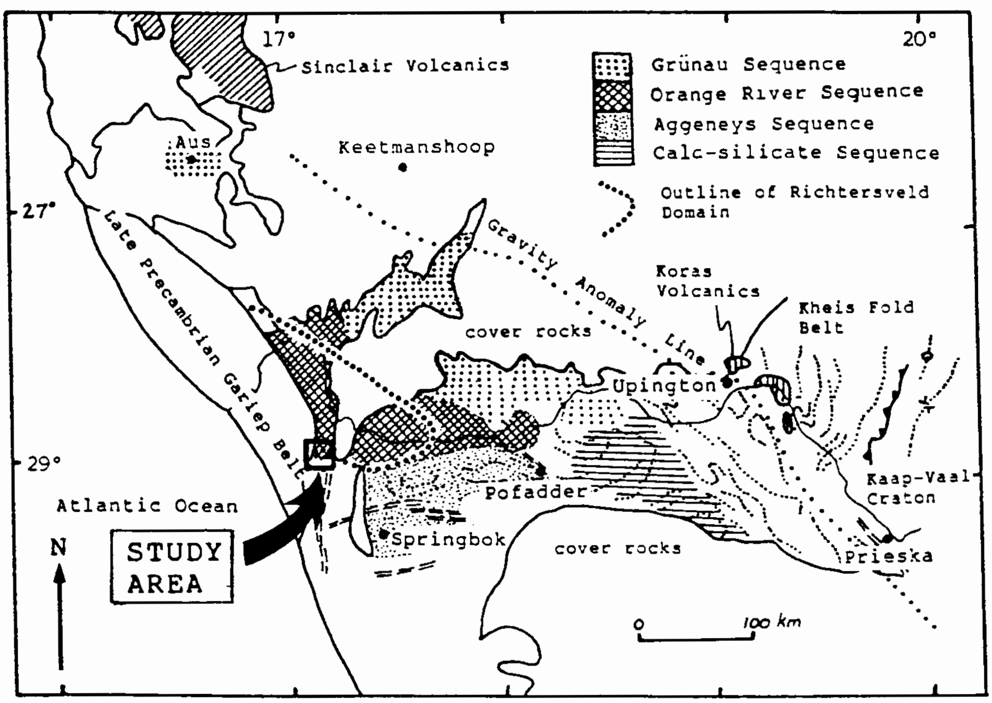


Fig. 2.2 The component parts of the Namaqua mobile belt (after Blignault et al., 1983, Fig. 2).

Blignault et al. (1983, p. 4) suggest that sequences of predominantly supracrustal lithotypes outline specific "domains" in the North-west Cape and southern Namibia. Here the Namaqua mobile belt is made up of four sequences, viz., the Grünau, Orange River, Aggeneys and Calc-silicate Sequences (Fig. 2.2). Stowe et al. (1983), Stowe (1984) and Joubert (1984) have modified their classification into subprovinces and terranes since recognising the fault-bounded nature of Blignault et al's "sequences" (Fig. 2.3). These authors interpret the Richtersveld and Namaqua lithologies to be temporally related and therefore to constitute one province, the Namaqua Province.

In the "Namaqua geotraverse", a narrow north-south strip between Vioolsdrif and Springbok (Fig. 2.3), major tectonic discontinuities such as shear zones or thrusts separate each terrane, with progressively greater uplift southwards from the Richtersveld. Thus the Okiep Terrane reveals an eroded crustal segment from the deepest part of the mobile belt (Blignault et al., 1983). Subdivision into terranes is based on stratigraphic, metamorphic, and structural characteristics which differ for each crustal segment (Table 2.2). These differences are borne out by radiometric data showing different background values for each terrane (average total counts per second), recorded from reconnaissance surveys between Vioolsdrif and Springbok (Table 2.2).

Geochemical characteristics of these terranes show that on the whole rocks of the Richtersveld Subprovince are more basic than those to the south, where high K_2O/Na_2O and Rb/Sr ratios are present (Reid, 1979b; Holland and Marais, 1983). Gravity surveys show that isostatic equilibrium has been reached over most of the north-western portion of the mobile belt, except for a steepening gravity profile in the extreme west near the coast (Muller and Smit, 1983).

2.3.2 Richtersveld Subprovince

The Richtersveld Subprovince straddles the Orange River; its broadest area of outcrop is in the west against cover rocks of the Gariep Group (Figs. 1 and 2.3). Rocks of the Vioolsdrif Terrane crop out in the westernmost part of this subprovince whereas those of the Pella terrane occur in the east (Fig. 2.3 and Table 2.1).

Lithologies of the Vioolsdrif Terrane consist mainly of extrusive and intrusive igneous types of calc-alkaline geochemistry, and metamorphosed to greenschist facies (Blignault, 1974; Ritter, 1980). Rb-Sr and U-Th-Pb isotopic ages indicate a 2000 -1900 Ma (Eburnian) history of calc-alkaline intrusive and extrusive I-type igneous rocks; protolithic material has been reworked and emplaced

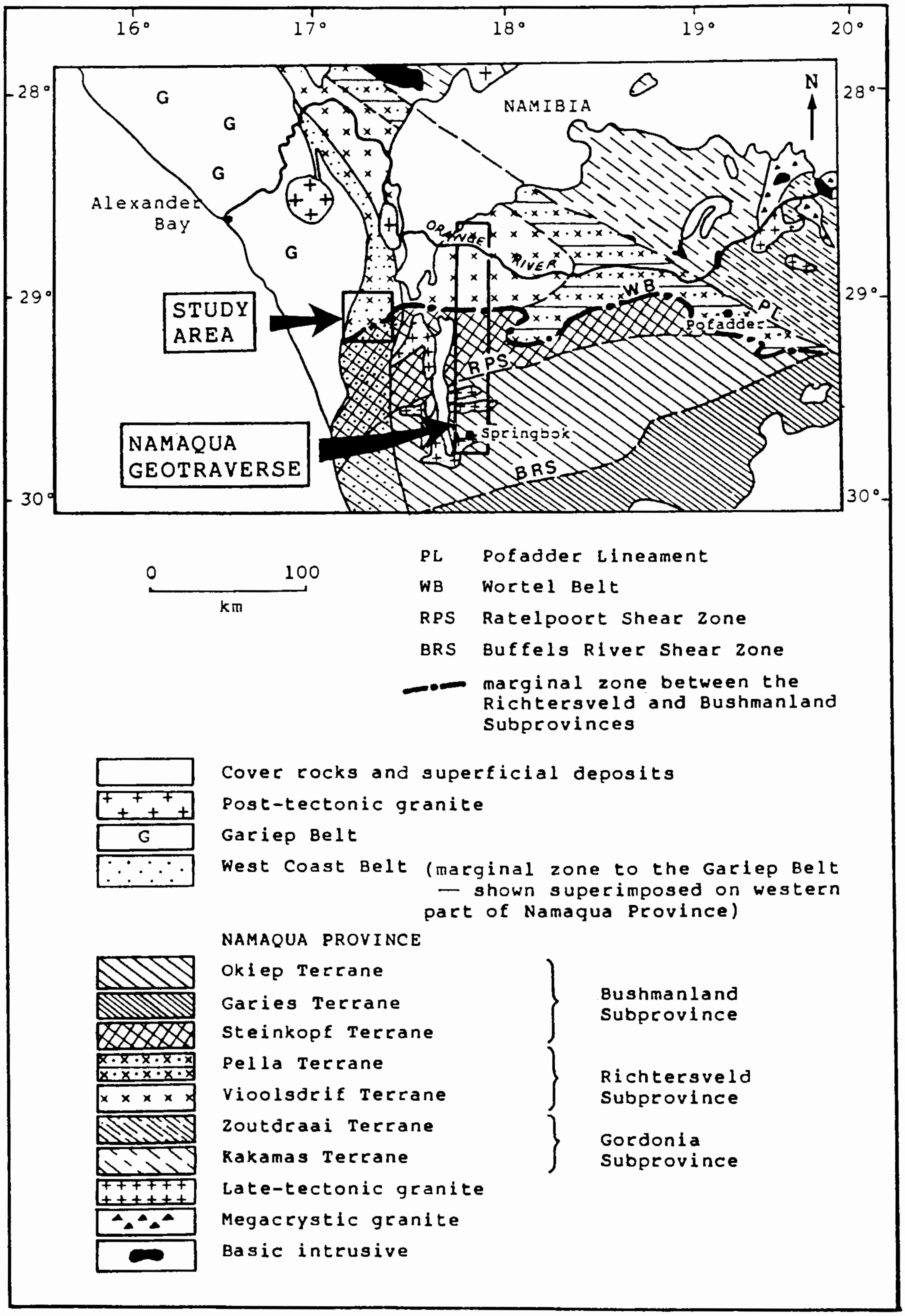


Fig. 2.3 Tectonic divisions, western Namaqua Province (portion of Fig. on p. 2B, Excursion Guide book, Conference on Middle to Late Proterozoic Lithosphere Evolution, Cape Town, 1984).

TABLE 2.2 Main characteristics of terranes in the Namaqua geotraverse (after Blignault et al. 1983; Muller and Smit, 1983.

Subprovince	Richtersveld	<----- Bushmanland	---->
Terrane	Vioolsdrif	Steinkopf	Okiep
Formations	Nous Tsams Windvlakte	Groothoek	Khurisberg
Lithologies	volcano-plutonic rocks	schists/ gneisses/ plutonic rocks	gneisses/ schists/ granulites plutonic rocks
Metamorphism	greenshist facies (2000 or 1200 Ma?)	amphibolite (1200 Ma)	granulite facies (1200 Ma)
Deformation	pre-1900 Ma	1800 and 1200 Ma	1200 Ma
Radiometrics i.e. (average total c.p.s.)	1250	1400	2100

as granitoid bodies at 1730 Ma (Reid, 1979b; 1981; Blignault, 1981; Reid and Barton, 1983). Tight to isoclinal folds in strata intruded by the 1900 Ma Vioolsdrif granodiorite bear evidence of an early (2000 Ma Eburnian) deformation event, prior to major batholithic intrusion (Blignault et al., 1983).

Low-grade (upper greenschist facies) supracrustal rocks of the Vioolsdrif Terrane (Nous and Tsams Formations; Orange River Group) crop out in a large area, extending eastwards from their contact with the Gariep Group for a maximum distance of some 100 km (Fig. 2.3, and Table 2.1). Farther eastwards (i.e. into Bushmanland) facies equivalents of these rocks (Guadom and Hom Formations of Orange River Group) are in medium grade (amphibolite facies) and constitute part of the Pella Terrane (Fig. 2.3 and Table 2.1; Blignault et al. 1983). Supracrustal rocks of the latter terrane comprise schists and metaquartzites (metasediments) of the Pella, Kabas, and Dabenoris Formations (Aggeneys Sequence) which overly, and are interpreted as younger than, gneisses and schists (metavolcanics) of the Guadom and Hom Formations of the Orange River Group (Colliston, 1983; and see Table 2.1). The Pella Terrane is therefore characterized by rocks of different metamorphic grade compared to the Vioolsdrif Terrane, but not separated from the latter by a fault boundary. This raises doubt as to whether the Pella Terrane does indeed constitute a separate terrane.

It is apparent that the Vioolsdrif and Pella Terranes constitute one crustal segment, the western part being in low grade whereas the eastern part has experienced medium grade metamorphism. However, an anomalous situation with respect to timing of the regional metamorphic imprint on the Vioolsdrif and Pella Terranes arises if current interpretations in this regard are taken into account. Ritter (1980) interprets metamorphism of the Vioolsdrif Terrane in the eastern Richtersveld as a 2000 - 1800 Ma(Eburnian) event which is related to intrusion of the Vioolsdrif Suite. Other researchers propose that regional metamorphism in the Namaqua mobile belt occurred during the 1200 - 1100 Ma Namaqua (Kibaran) event (Clifford et al., 1974, 1981; Joubert, 1981; Blignault et al., 1983; Moore, 1986; see Table 2.1). This is based on interpretations that supracrustal rocks in the Pella and Steinkopf Terranes were deformed during the latter event. The effect of the Namaqua metamorphic event on the Vioolsdrif Terrane still remains unresolved.

In Bushmanland the boundary between the Pella Terrane of the Richtersveld Subprovince and that of the Steinkopf Terrane(Bushmanland Subprovince) is denoted as the Wortel Belt (Fig. 2.3). The western extension of this boundary to the Gariep Group is referred to in this thesis as the Groothoek Thrust zone. The relationship of the Pella Terrane to the Steinkopf Terrane does not form part of the present investigation because the former terrane is not present in the study area, but lies about 100 km farther east. However, this relationship is significant in a regional context, and will be examined in a synthesis of the deformational and metamorphic history of the region, when geotectonic models are considered (Chapter 6).

The nature of the boundary between the Richtersveld and Bushmanland Subprovinces remains problematical because:
- rock types of the Richtersveld Subprovince may be present to the south of the Vioolsdrif Terrane. In the high-grade Okiep Terrane, 60 to 100 km to the south of the Vioolsdrif Terrane, Holland and Marais (1983) interpret rock types such as the Lammerhoek gneiss as correlates of the Vioolsdrif Suite. A similar situation is proposed by these investigators in the area 30 km east of Port Nolloth where rocks resembling those of the Vioolsdrif Suite in field appearance intrude the Een Riet Subgroup (Steinkopf Terrane). If this can be confirmed through isotopic data, then rocks of the Vioolsdrif Suite have a much greater southward distribution than

has been recognized in the past. This is a fundamentally important factor in understanding the relationship between lithologies of the Steinkopf/Okiep Terranes and those of the Vioolsdrif Terrane;
- there are different interpretations as to when the foliation developed in gneisses of the Richtersveld and Bushmanland Subprovinces. Kröner (1975) draws attention to the possibility that some of the gneisses of the Pella Terrane may be reconstituted rocks of the Vioolsdrif Terrane. Near Goodhouse, just east of the Richtersveld, Kröner et al. (1983) interpret the age of this foliation as older than 1862 Ma, based on the age of the Goodhouse granite which transects the foliation there. A different interpretation is put forward by Barton (1983) who regards the main penetrative foliation in the Namaqua mobile belt as forming during the Namaqua event (1200 - 1100 Ma).

In the Pella area north of Pofadder, Colliston et al. (1981) and Blignault et al. (1983) show that volcanic rocks of the 2000 Ma Guadom Formation (Orange River Group) can be traced from the Vioolsdrif Terrane eastwards along strike into the Pella Terrane, where they are foliated and gneissic as a result of overprinting during the Namaqua event.

In the Namaqua geotraverse Blignault et al. (1983) propose that the foliation there is modified by the Groothoek Thrust, a major tectonic discontinuity separating the Richtersveld and Bushmanland Subprovinces north of Steinkopf. Thrust movement is regarded by these authors (op. cit.) as taking place mainly during syn- Spektakel Suite (1100 Ma) times.

2.3.3 Bushmanland Subprovince

2.3.3.1 Steinkopf Terrane

This terrane, occurring immediately south of the Richtersveld Subprovince, has a north-south width of about 60 km, and consists of medium-grade (amphibolite facies) rocks (Fig. 2.3; and Table 2.1). In the central portion of the Namaqua geotraverse Van Aswegen (1983; 1988) records some relict granulites.

The Steinkopf Terrane is characterized mainly by three silicic orthogneiss units (Gladkop Suite), and metasediments (Een Riet Sub group; see Table 2.1). The oldest orthogneiss of the Gladkop Suite, the Steinkopf gneiss, intrudes metasediments of the Een Riet Subgroup. The Brandewynsbank and Noenoemaasberg orthogneisses are younger than the Steinkopf gneiss (Van Aswegen, 1983). Blignault et al. (1983, p. 13) correlate the 1800 Ma Een Riet Subgroup with the Aggeneys Sequence, whereas field evidence in the Pella area reveals that the latter is younger than the Guadom Formation (Orange River Group), i.e. less than 2000 Ma (Colliston, 1983). Thus, metasediments of the Steinkopf Terrane are interpreted through field evidence as younger than lithologies of the Richtersveld Subprovince.

Van Aswegen (1983), however, finds that metasediments of the Steinkopf Terrane are intruded by Vioolsdrif-like rocks. This suggests that metasediments of the Steinkopf Terrane are older than lithologies of the Vioolsdrif Terrane. This paradox can only really be resolved through systematic isotopic dating of appropriate lithologies of both terranes.

Rb-Sr and Pb-Pb dating methods give an age of 1800 Ma for the Gladkop Suite (Barton et al., 1981), but Van Aswegen (1983) suggests that emplacement of the suite was probably contemporaneous with that of the high level Vioolsdrif Suite batholiths of the Vioolsdrif Terrane to the north. Subsequent deformation took place between the 2000 Ma Eburnian and 1200 Ma Namaqua orogenies (Barton et al., 1981), suggesting that the gneisses of the Gladkop Suite acquired their tectonic fabric at about 1800 Ma. Tectonic reworking of this fabric occurred during the 1200-1000 Ma Namaqua event (Blignault et al., 1983, p. 17). Distribution of the Gladkop Suite west of the area investigated by Van Aswegen (1983; 1988) is as yet unknown, and needs to be traced on a regional basis.

The Konkyp augen gneiss, which intrudes the Een Riet Subgroup, presents a problem regarding its age and position in the stratigraphic sequence (Reid and Barton, 1983). Rb/Sr isochrons yield an age corresponding to the Namaqua event, suggesting that it could be correlated with the Little Namaqualand Suite. Based on field evidence, however, Reid and Barton (1983) consider that it could represent an 1800 Ma granitoid.

2.3.3.2 Okiep Terrane.

This crustal segment is characterized by a preponderance of granitic intrusives, and high-grade granulite facies rocks formed in the temperature range 800°-900°C and pressures of 6-7 kbar (Clifford et al., 1981). It contains cupriferous noritoid bodies centered around Okiep. Augen gneisses of the Little Namaqualand Suite (Nababeep gneiss, Modderfontein gneiss) are widespread; they are dated at 1200 Ma (Clifford et al., 1974). Post-tectonic Concordia, Rietberg and Kweekfontein granites of the Spektakel Suite are intrusive into metasediments and metavolcanics of the Khurisberg Formation, and are dated at 1100 Ma (Clifford et al., 1974; Marais, 1981).

In this crustal segment deformation events are virtually impossible to reconstruct because of the very high proportion of granites compared to supracrustals. Clifford et al. (1974), however, propose a structural model of early isoclinal folding with concomitant axial planar foliation, but this interpretation is not recognized by geologists of the Okiep Copper Company (P. Joubert, personal communication, 1984).

According to Holland and Marais (1983, p. 84) it is possible that rocks of the Vioolsdrif Suite are present in the Okiep Terrane. Based on this proposal Colliston and Praekelt (1988) position the boundary between the Richtersveld and Bushmanland Subprovinces as somewhere within the Okiep Terrane. As mentioned in Section 2.2.2, conclusive proof of absolute age relationships between lithologies of the Okiep, Steinkopf and Vioolsdrif Terranes is necessary before the latter interpretation can be accepted, or before one can justify using the term "terrane".

Confusion is rife with regard to defining tectonic subdivisions of the Namaqua Province. This confusion comes about (i) because researchers apply various terms such as sequence, province and terrane to mean the same thing; and (ii) there is not enough isotopic data available to precisely define terranes of the Namaqua Province.

2.4 AGE RELATIONSHIPS

2.4.1 Introduction

From discussions in section 2.3 it is apparent that age relationships between rocks of the Richtersveld and Bushmanland Subprovinces have an important bearing on the positioning of the contact between the two subprovinces. Although isotopic data are available there is conjecture regarding the ages of certain rocks, and field relationships.

It is apparent that most interpretations of the regional geological framework discussed in section 2.3 propose the juxtaposition of the 2000 Ma Vioolsdrif and Pella Terranes in the north across those of the assumed 1800 Ma Steinkopf Terrane in the south. This appraisal implies an overthrust relationship along the zone separating the Steinkopf from the Vioolsdrif and Pella Terranes. The Groothoek Thrust is interpreted as a tectonic discontinuity separating the Vioolsdrif from the Steinkopf Terrane, just north of Steinkopf (Blignault et al., 1983). Conversely, if the suggestion that rocks of the Bushmanland Subprovince can be proved to be older than those of the Richtersveld Subprovince then the positioning of the boundary between the two subprovinces will have to be interpreted in a different light.

A complicating factor in medium to high-grade metamorphic terranes is that the law of superposition cannot be applied to interpret facing direction of strata, as primary structures are usually obliterated. This is the case in the study area and poses a fundamental problem on the timing and extent of the metamorphic imprint which traverses the Vioolsdrif and Steinkopf Terranes. Is this imprint related to the Eburnian event as proposed by Ritter (1980), or is it related to the Namaqua event, as proposed by Blignault et al. (1983)?

The interpretation of isotopic ages in relation to the tectonic and metamorphic history therefore becomes vitally important in resolving the problem of the relationship between rocks of the Richtersveld and Bushmanland Subprovinces.

2.4.2 Field evidence vs. isotopic evidence

Extrusive igneous rocks (Orange River Group) and co-genetic intrusives (Vioolsdrif Suite) are accepted as having formed between 2000 - 1900 Ma (Reid, 1979b), even though the base of the former is nowhere exposed and relationship to basement of the entire rock sequence is therefore unknown. A younger more acidic phase intruded at about 1730 Ma (De Villiers and Burger, 1967; Reid, 1979a and b; 1981; SACS, 1980; Reid and Barton, 1983).

It is important to determine whether the southern contact of lithologies of the Vioolsdrif Terrane with those of the Steinkopf Terrane is a normal superposition of strata, or a major tectonic discontinuity. On field evidence Ritter (1978; 1980) interprets this as a normal transition, i.e. rocks of the Richtersveld Subprovince stratigraphically overly those of the Bushmanland Subprovince. Subsequent metamorphism took place between 2000 - 1800 Ma. This interpretation contrasts with that of Blignault et al. (1983), and Colliston (1983) who regard rocks of the Bushmanland Subprovince as younger than those of the Richtersveld Subprovince.

Isotopic age determinations on rocks of the Aggeneys Sequence range between 1600 Ma and 1200 Ma (Köppel, 1980; Reid, 1981; Betton, 1984), thus confirming Colliston's field observations (section 2.3.2).

Clifford et al. (1981) maintain that segments of the high-grade metamorphic sequence in the Springbok area can be correlated with the quartzite-schist succession of the Aggeneys Sequence at Aggeneys and Gamsberg, near Pofadder.

Similarly, on lithological grounds Blignault et al. (1983) correlate the Aggeneys Sequence with the Een Riet Subgroup farther west in the Steinkopf Terrane. This is a rather long range correlation, bearing in mind that Colliston et al. (1981) note that metasedimentary and metavolcanic rocks at Geselskapbank to the west of Pofadder cannot satisfactorily be correlated with any portion of the Aggeneys Sequence. The structural fabric in the Een Riet Subgroup suggests that their age may be closer to 2000 Ma (Blignault et al., 1983, p. 6). A Rb/Sr age of 950 Ma for metasediments of the Een Riet Subgroup obtained by Barton (1983, p. 53) is, however, interpreted as a minimum age for these rocks.

The youngest Spektakel Suite granites, viz. the Concordia, Rietberg and Kweekfontein granites, dated at ≈1100 Ma (Blignault et al., 1983, p. 26), represent products of mantle-derived potassic syenitoid melts (Barton, 1983; Reid and Barton, 1983). Intrusive augen gneisses of the Little Namaqualand Suite (Nababeep and Modderfontein gneisses) reflect a main granulite facies metamorphic age of ≈1200 Ma in the Okiep region (Clifford et al., 1974). Granites older than the above, but still incorporated in the Spektakel Suite, include the Eyams granite, a ≈1500 Ma leucogranite in the west of the Steinkopf Terrane, and the Konkyp granite in the north of this terrane (Barton, 1983). Although the Konkyp granite intrudes the Een Riet Subgroup metasediments it contains the Steinkopf fabric (1800 Ma age?), and its inclusion in the Spektakel Suite is therefore not justified (Reid and Barton 1983).

In conclusion, isotopic evidence suggests that rocks of the Richtersveld Subprovince are older than those of Bushmanland Subprovince. Field evidence corroborates this in the Pella area (Colliston, 1983). In the southern Richtersveld region, however, rocks of the Richtersveld Subprovince overlie those of the Bushmanland Subprovince, and are therefore considered younger than the latter (Ritter, 1980). Quite clearly the Aggeneys Sequence cannot be correlated as yet with the Een Riet Subgroup because isotopic and field evidence are contradictory. Rocks resembling those of the Vioolsdrif Suite in field appearance occur in the Okiep Terrane, but have not yet been positively correlated with the Vioolsdrif Suite (Holland and Marais, 1983). The conclusions reached above reveal that major regional correlation problems exist when examining the relationship between the Richtersveld and Bushmanland Subprovinces.

2.5 TECTONIC MODELS

2.5.1 Early to Mid-Proterozoic Tectonism

2.5.1.1 Cordilleran Model

Several authors invoke the Cordilleran tectonic model to account for observed geologic features and geochemical characteristics in the north-western portion of the Namaqua mobile belt (Reid and Barton, 1983; Barton, 1984; Reid, 1984). De Beer and Meyer (1983) regard this model as appropriate for explaining gravity data across the northern and north-eastern mobile belt/Kaapvaal craton boundary.

Based on a regional mapping project, Ritter (1983) concludes that a diapiric model explains all the structural, metamorphic and magmatic complexities in the eastern Richtersveld region. Magmatism, deformation and metamorphism are envisaged as mutually interdependent processes, intensities of deformation revealing different crustal levels.

Holland and Marais (1983) favour a model of granite under-plating of sialic crust, with concomitant sedimentation above. This model is proposed on the basis of geochemical evidence in the Okiep Terrane (Holland and Marais, 1983). Tectonic reworking is visualized as taking place along linear belts within one large ensialic plate. This idea is similar to that of Kröner (1977a) who suggests that mobile belts in Africa do not originate through large-scale movement of plates in the Proterozoic (i.e. cratonization). Instead he proposes that they are created as long linear belts by mantle plume upwelling which weakens large ensialic continental plates. This mechanism is thought to have been in existence since the end of Archaean times.

Moore (1984; 1986) proposes that the early structural setting of paragneisses of the Bushmanland Subprovince is one of shallow water deposition in a marginal rift environment. This setting is located across the 2000 Ma volcano-plutonic rocks of the Richtersveld Subprovince to the north, and continentally derived sediments of the Bushmanland Subprovince to the south.

2.5.1.2 Continent - Continent Collision

Coward (1980a) suggests that a continental collision model with thrusts developing southwards can explain the large volume of granitic material associated with flat-lying gneisses in Namaqualand. De Beer and Meyer (1983), and Van Zijl et al. (1984), however, discard a continental collision model, as they find that geophysical surveys in Namaqualand and Bushmanland to date have not located any zones which could be interpreted as ophiolitic.

Joubert (1986a) proposes an accretion model for the Namaqua Province as a whole, suggesting that microplates were accreted during a 2000 Ma event, and a 1300-1200 Ma period. The suture zones of micro-continents are demarcated by prominent zones of mafic and ultramafic rocks.

2.5.1.3 Over-thrust Model

Based on evidence from the Namaqua geotraverse, Blignault et al. (1983) and Van der Merwe and Botha (1989) propose an over-thrust model whereby major southward-developing thrust zones form over a north-south distance of some 200 km (Fig. 2.4). The emplacement of large thrust sheets could explain some of the metamorphic imprints observed in this area. Of interest locally is that the aforementioned authors interpret the Groothoek Thrust, which

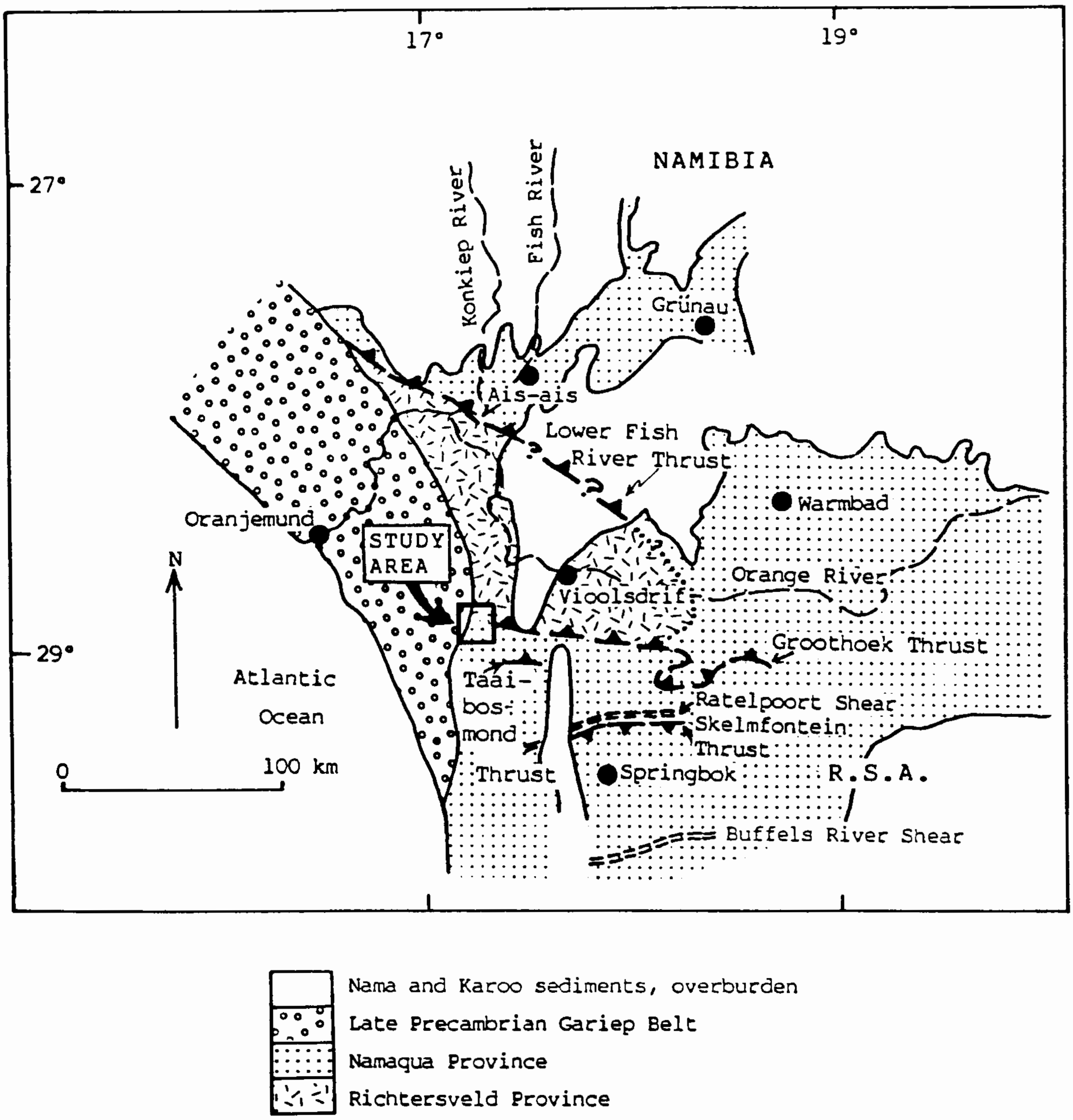


Fig. 2.4 The Regional geology of the north west Namaqua Province showing location of major thrust/shear zones (modified slightly after Van der Merwe and Botha (1989).

separates the Richtersveld from Bushmanland Subprovinces, as forming at the same time as the Spektakel Suite. It shows an imprint developed in response to protracted tectonism during the Namaqua event (Blignault et al., 1983). This contrasts with the ideas of Moore (1986) who proposes an earlier age (≈1800 Ma) for the origin of the Groothoek Thrust. The Taaibosmond and Skelmfontein Thrusts lie south of the Groothoek Thrust, the latter approximately at the tectonic discontinuity separating the Steinkopf and Okiep Terranes (Van der Merwe, 1986).

2.5.2 Late Proterozoic/Early Palaeozoic (Pan-African) Tectonism and Metamorphism

Joubert (1971) drew attention to north-northeast striking shear zones in the coastal belt of Namaqualand, interpreting a left-lateral sense of movement associated with this tectonism. Ritter (1978, p. 96) concluded that strong left-lateral shearing took place between 750-1000 Ma in the south-east Richtersveld. The latter author interprets structures associated with the northern extremity of the Steenbok Shear as due to compression.

Kröner (1974) proposed a plate tectonic model to explain Late Proterozoic tectonism of the Gariep Belt. The model incorporates an initial east-west directed rifting phase with predominant vertical tectonism, followed by easterly-directed compression.

Allsopp et al. (1979) observed that zones of brittle deformation in the basement, associated with Pan-African tectonism, increase west-wards. This tectonism coincides with resetting events at 700 and 500 Ma.

In the area south of the Gariep Belt, along the west coast, Joubert and Waters (1980) and Waters, Joubert and Moore (1983) have interpreted a lower amphibolite facies metamorphic imprint on the gneisses of the Bushmanland Subprovince as younger than the Namaqua event, but older than Nama sedimentation (i.e. >600 Ma). Jackson and Zelt (1984) relate this metamorphic event to the extensional phase of Gariep tectonism.

The present investigation poses the question, What tectonic and metamorphic effects does the Pan-African event have on the western extremity of the zone separating the Richtersveld and Bushmanland Subprovinces, and can the history be deciphered?.

2.6 SUMMARY

2.6.1 Regional metamorphic events

The main metamorphic events recorded in the western portion of the Namaqua Province are:

- a greenschist facies imprint in low grade rocks of the Richtersveld Subprovince, interpreted as having formed during a 2000-1800 Ma event (Ritter, 1980; see Table 2.1);
- an amphibolite to granulite facies imprint recorded in the major outcorp area of the Namaqua Province, formed between 1200-1100 Ma (Clifford et al., 1974);
- a metamorphic imprint at 700 Ma and another at 500 Ma are recorded in the marginal zone to the Gariep Belt (i.e. West Coast Belt). The former reached lower amphibolite facies in southern Namaqualand in the Bitterfontein area (Waters, Joubert and Moore, 1983) whereas greenschist facies is associated with the latter.

2.6.2 Relationship between the Richtersveld and Bushmanland Subprovinces

The first problem concerns the manner in which metasedimentary, metavolcanic, and intrusive rocks of the Richtersveld and Bushmanland Subprovinces relate to each other. Field evidence in the south-eastern Richtersveld reveals that rocks of the Richtersveld Subprovince overlie those of the Bushmanland Subprovince i.e. the latter are interpreted as older (Ritter, 1980), yet isotopic evidence from suites of the Bushmanland Subprovince shows that all ages are younger than the 2000 Ma Orange River Group. This anomaly remains unresolved if the Een Riet Subgroup in the Steinkopf area is correlated with the Aggeneys Sequence as proposed by Blignault et al. (1983), because this implies that the former is younger than the Orange River Group (cf. Colliston, 1983). The problem focuses on the validity of correlating the Een Riet Subgroup with the Aggeneys Sequence, and why Ritter's stratigraphic interpretations are not consistent with isotopic evidence.

In the zone separating the Richtersveld and Bushmanland Subprovinces all rocks have such a strong tectonic and metamorphic overprint that contact relationships are not clear. This makes lithological relationships difficult to decipher. Can rocks of the Vioolsdrif Suite be positively identified south of the type area in the region of overprinting, as proposed by Holland and Marais (1983), and if so, how far south of this zone do they occur?

A further question brought to light through the literature search concerns the age of the foliation in the zone separating the Richtersveld and Bushmanland Subprovinces. Kröner et al. (1983) interpret it as older than the 1862 Ma Goodhouse granite which transects the foliation. Barton (1983), however, regards its development as due to overprinting during the Namaqua event (i.e. ≈1200 Ma).

The age of the Konkyp granite remains an unresolved problem, a minimum age of 950 Ma being recorded, but it may be an 1800 Ma Vioolsdrif granitoid (Reid and Barton, 1983).

2.6.3 Groothoek Thrust

In the Namaqua geotraverse, Van der Merwe (1979), Theart (1980, p. 78) and Van der Merwe and Botha (1989) propose a thrust zone in the Groothoek schists some 20 km north of Steinkopf to account for the strongly developed fabric in these rocks. In the area just north and west of Pofadder the position of the Groothoek Thrust appears to be related to mafic/ultramafic rocks and associated magnesian schists (Joubert, 1986a).

North of Steinkopf (in the Namaqua geotraverse) Blignault et al. (1983) position the Groothoek Thrust between lithologies of the Richtersveld and Bushmanland Subprovinces, but delineate it across an area of overburden. A similar situation is present in the area immediately east of the geotraverse (Strydom, 1982). Associated with the thrust is a zone of 'muscovitization' and retrograde metamorphism. The thrust is considered by these authors (op.cit.) to have developed during Spektakel Suite times, thus assigning it an age of ≈1100 Ma, with protracted movement related to the Namaqua event. In contrast Moore (1986) suggests that the thrust developed during an earlier collision-related event.

Four questions are raised.

(i) How far west of the Namaqua geotraverse can the Groothoek Thrust be traced?

(ii) If it does persist westwards, can the thrust zone be demarcated in basement rocks of the West Coast Belt where Late Proterozoic tectonism, intrusions and metamorphism have greatly modified and complicated the earlier history of the area?

(iii) Does Blignault et al.'s (1983) model conform with the metamorphic history of the area?

(iv) Can the timing of Groothoek thrusting be resolved?

2.6.4 Pan-African Tectonism and Metamorphism

Most investigators who have worked in the West Coast Belt interpret northerly-striking Pan-African shear zones in this region as having a left-lateral sense of movement. Several episodes of deformation are associated with these shears, but their history still has to be unravelled. Late-Proterozoic to Early Palaeozoic metamorphism associated with the shearing event is not clearly understood.

2.6.5 Geodynamic Models for the Early to Mid-Proterozoic

A variety of tectonic models are currently invoked to explain deformation, magmatic and metamorphic events associated with the Early to Mid-Proterozoic history of the Namaqua Province. These include Cordilleran, continental collision and over-thrust models. Some of these models have application on a local scale, but not necessarily on a macro-scale. These models are not necessarily mutually exclusive.

In the study area unraveling the Early to Mid-Proterozoic history of the Namaqua Province is complicated by tectonic and metamorphic overprinting during the Pan-African event. Tectonic models which could satisfactorily explain the complex history of this region are currently lacking.

2.7 PROBLEMS DEFINED THROUGH GEOLOGIC MAPPING IN AN AREA SOUTH-EAST OF EKSTEENFONTEIN

During the present investigation an area traversing the boundary between the Richtersveld and Bushmanland Subprovinces was mapped on a scale of 1:36,000 (Appendix A-1). It extends from the eastern margin of the Gariep Belt eastwards to approximately longitude 17° 30′ (Annexure 1). Geologic anomalies arising from this survey, and problems involving regional stratigraphic correlation with lithologies of the Namaqua geotraverse, are presented below.

2.7.1 Variation in foliation trend

A north-northeasterly striking (Pan-African) foliation is superimposed across that of the northwesterly-striking trend in basement rocks as steeply-dipping shear zones, averaging 4 km apart. These shear zones are named from west to east, as follows: Stinkfontein Contact Shears, Kromnek, Tierkloof, Steenbok (Joubert, 1971), Kouefontein, Riethoek, Chabiesies and Witbakensberg Shears (Annexure 1).

Foliation trends older than Pan-African are best developed in metavolcanics and metasedimentary rocks. For the most part these trends are consistently towards the north-west, but there are local variations within the 18 km broad north-south zone surveyed. In the extreme north-east of this area the predominant foliation strikes north-west, but just to the south a consistent east-west strike is discernible, defining lens-shaped units in the supracrustal formations. In the southern portion of the area this older foliation trend is variable, but in places is notably discordant to that farther north. These divergences in foliation trend demand explanation.

2.7.2 Anomalous Distribution of Metasedimentary/ Metavolcanic Rocks

2.7.2.1 Metapelitic/psammitic rocks

Lithologies of the Groothoek suite show marked differences in outcrop width along strike, especially between the Steenbok and Tierkloof Shears, where outcrop is broader compared to the equivalent stratigraphic unit east and west of these shears (Fig. 2.5 and Annexure 1). Similarly, between these two shears there is only a small outcrop of Groenrivier suite rocks (overlying Groothoek suite), compared to the areas east and west of these two shears (Annexure 1). Vioolsdrif Granodiorite is in contact mainly with the Groothoek suite here, whereas to the west and east of these shears it is in contact with metavolcanics of the Groenrivier suite. An explanation needs to be found to account for the anomalous outcrop width of the two formations.

In the south of the area gneissic and schistose rocks are allocated to the Chabiesies suite. The stratigraphic position of these rocks in relation to the Steinkopf Terrane needs to be verified.

2.7.2.2 Metavolcanic rocks of the Windvlakte suite

The upper metavolcanic unit of the Windvlakte suite crops out in the north of the area mapped and is located only in two areas, (i) between the Steenbok and Tierkloof Shears, and (ii) west of the Kromnek Shear. In area (i) the Vioolsdrif granodiorite separates them from schists and gneisses of the Groothoek suite to the south. In area (ii) they overlie the Tweeriviere granite gneiss of the Vioolsdrif Suite (Annexure 1). Their relationship to the lower unit of the Windvlakte suite is unknown and needs to be investigated, as does the cause of their restricted geographical distribution.

2.7.2.3 Quartzites of the Ratelfontein suite

Quartzites occurring in the Ratelfontein suite (cf. Fig. 2.5), unlike other quartzite units mapped in this area, are characterized by separate boudin or "pip" shapes. The larger-sized boudins occur in the extreme east, but generally become progressively smaller westwards along strike. Interpretations of their distinctive shapes, and occurrence especially along the base of this stratigraphic unit need to be advanced.

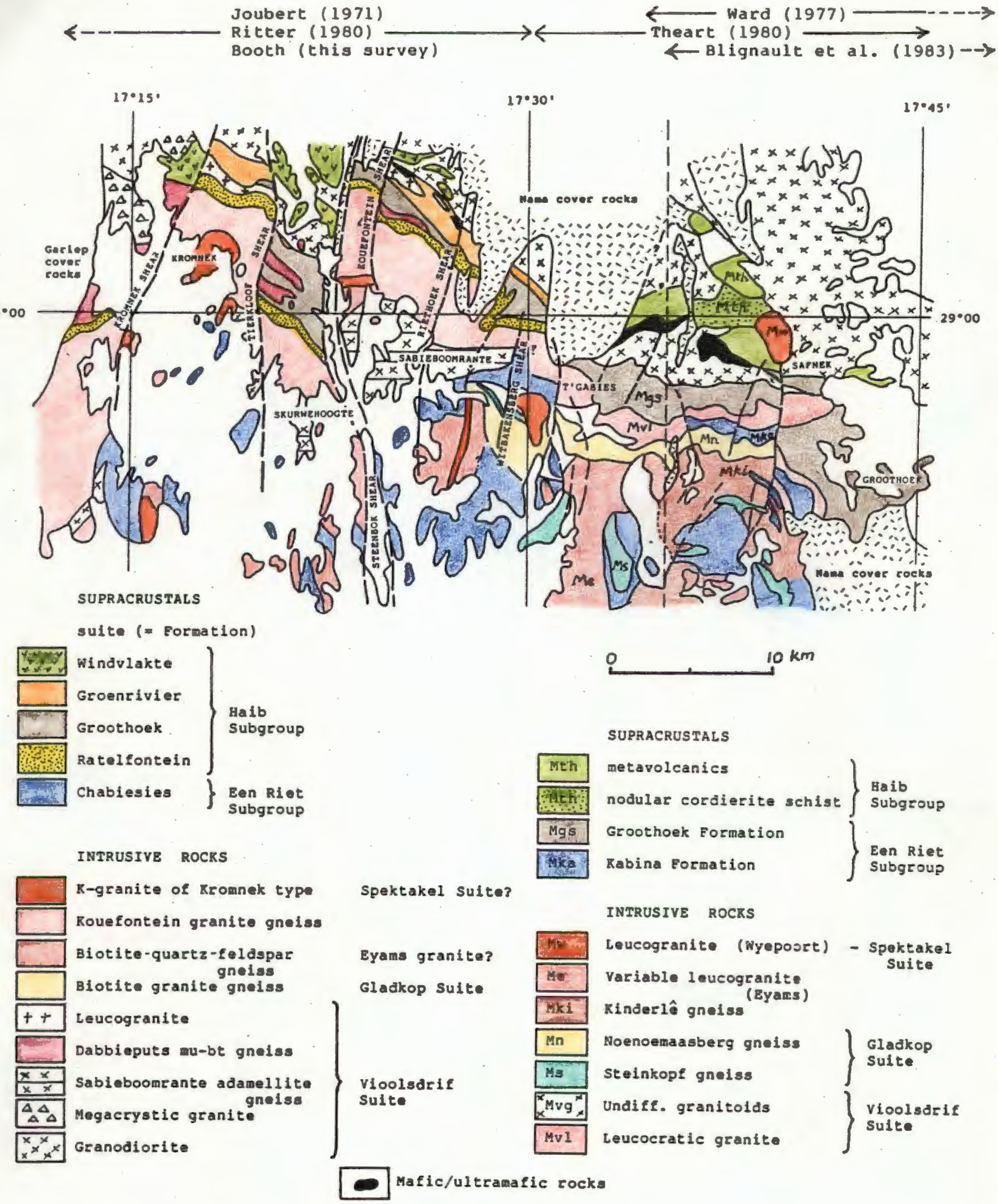


Fig. 2.5 Regional geological map showing correlation of lithologies from study area (west of 17°30') with those of the Namaqua geotraverse (east of 17°30'). Information based on:
(i) West of 17°30' - Joubert (1971); Ritter (1980); Booth (this survey);
(ii) East of 17°30' - Ward (1977); Theart (1980); Blignault et al. (1983).

2.7.2.4 Distribution of the Nama Group

In the area of investigation Nama Group sediments are confined to the area east of the Steenbok Shear, occurring as outliers at different elevations, and a linear slice within the Steenbok Shear. An explanation for their abrupt termination and anomalous disposition east of the Steenbok Shear is sought.

2.7.3 Problems Relating to Igneous Intrusive Rocks

2.7.3.1 Sabieboomrante adamellite gneiss

This rock occurs in the south-eastern portion of the study area, east of Steenbok Shear (Annexure 1). It has a triangular outcrop shape, the apex pointing eastwards. Its northern contact is concordant with the strata whereas its southern contact is transgressive (Annexure 1). In field appearance the rock resembles Vioolsdrif Granodiorite to the north, but it has an augen texture and is strongly foliated. Four questions arise from the field observations.

(i) Is this an intrusive rock, or is it a sheared end-product of metasediments and/or metavolcanics, as interpreted by De Villiers and Söhnge (1959, p. 108)?

(ii) What is the contact relationship between this rock and its neighbours?

(iii) What is the age of this rock?

(iv) Is the foliation in this rock an expression of the 1200 Ma Namaqua event, similar to that of Little Namaqualand Suite augen gneisses in the Okiep area, or is it an older foliation ($\approx$1800 Ma) similar to that described by Kröner et al. (1983) in the Goodhouse area, and south of Vioolsdrif by Ward (1974)?

If the Sabieboomrante adamellite gneiss is intrusive and a correlate of Vioolsdrif Granodiorite then its markedly smaller size and form of intrusion, compared to Vioolsdrif Suite batholiths, need explanation. The augen texture in Sabieboomrante adamellite gneiss is not characteristic of Vioolsdrif Granodiorite in the type area.

In the south of the area, west of the Steenbok Shear, granitoid bodies similar in appearance to the Sabieboomrante adamellite gneiss occur at Skurwehoogte (grid reference K 13; Annexure 1), and just west of the Tierkloof Shear (grid reference I 9; Annexure 1). Similar bodies crop out sporadically for some 20 km southwest of the present area where they occur as strongly foliated rocks on the farm Oograbies. Are these rocks related to Sabie-boomrante adamellite gneiss east of Steenbok Shear, and Vioolsdrif Granodiorite farther north ?.

2.7.3.2 Megacrystic Granite

A megacrystic, dark-coloured granite occurs near the contact with the Stinkfontein Formation in the extreme north-west of the area. The problem here concerns what relationship this rock has to Vioolsdrif Granodiorite. It resembles the latter along at least one contact with it, but the rock has a tectonised fabric over the largest part of its area of outcrop. Its relationships with its neighbours is for the most part obscured by tectonism which took place during the Pan-African event.

2.7.3.3 Kouefontein granite gneiss

This rock is described as "muscovitized granite" (De Villiers and Söhnge, 1959), "aplogranite gneiss" (Joubert, 1971), and is part of the "Pink Gneiss Unit" in the south-eastern Richtersveld (Ritter, 1980). In the area of present study it forms a thick, concordant sheet separating metasediments in the north from those in the south, and makes up the north-westerly trending watershed.

Are these rocks intrusive? There is conjecture over the origin of similar rocks in other parts of Namaqualand, where some pink gneisses are shown to be of sedimentary origin (Lipson, 1980; Duncan et al., 1984b).

Blignault et al. (1983) allocate this rock to the Vioolsdrif Suite where it appears in the Namaqua geotraverse as a thin concordant unit (code Mvl in Fig. 2.5). However, correlation of this leuco-granite with lithologies in the area of present study, and classification into either the Vioolsdrif, Little Namaqualand or Spektakel Suites, needs verification. Its occurrence at the junction separating two significantly different lithological and structural trends in supracrustals may have an important bearing on the interpretation of Mid-Proterozoic tectonism in the north-western part of the Namaqua mobile belt.

2.7.3.4 Mafic/Ultramafic Rocks

In the north-east of the study area coarse-grained serpentinites and associated amphibolites form a linear outcrop over a strike length of 7 km, but less than 300 m wide (Annexure 2). Serpentinites also occur some 4 km farther south, within the Kouefontein granite gneiss, but form only three small oval to circular, widely-spaced entities.

Three questions arise concerning the nature of these rocks.

(i) Are these rocks intrusive into meta-volcanics/ granodiorites in the north-east and pink gneisses in the central portion of the area, or are they xenoliths?

(ii) Do linear outcrops of mafic/ ultramafic rocks in the north-east represent tectonically emplaced slivers of ocean floor rocks?

If so, then do the isolated occurrences in Kouefontein granite gneiss farther south represent separate intrusions?

(iii) Is the age of these rocks the same as that for mafic rocks occurring farther eastwards along strike in the Wortel Belt near Pofadder?

A Pb-Pb isochron yields a 2200 Ma age for the latter mafic intrusives (Welke and Smith, 1984, quoted in Excursion Guidebook, Conference on Middle- to Late- Proterozoic Lithosphere Evolution, Excursion B). One of the mafic bodies in the southern Richtersveld is dated at 1858 ± 223 Ma (Reid 1979a), which is similar to the 1990 ± 125 Ma age for a mafic body on the farm, Wortel, near Pofadder (Welke and Smith, 1984).

2.7.3.5 K-rich Granites of Kromnek type

Pinkish to grey granite intrusives, mostly small sized plutons, are located mainly south of the Ratelfontein suite/ Kouefontein granite gneiss contact. Their age needs to be determined to find out if they can be correlated with the 1100 Ma Spektakel Suite, the 920 Ma Richtersveld Suite, or whether they represent a separate suite.

Does their spatial distribution mainly south of the Ratelfontein suite bear some relevance to a particular geodynamic model?

2.7.4 Problems Relating to Tectonism and Metamorphism

2.7.4.1 Timing of deformational events

Field observations and petrographic studies of deformation textures reveal evidence of multiple deformation and metamorphic events. Restoration of tectonic events associated with Late-Proterozoic/Early Palaeozoic (Pan-African) tectonism is necessary before a clearer understanding of Early to Mid-Proterozoic tectonism can be gained.

2.7.4.2 Early to Mid-Proterozoic and Late Proterozoic to Early Palaeozoic (Pan-African) metamorphic imprints.

From an assessment of the regional metamorphic zonation pattern in Namaqualand and the Richtersveld, it is apparent that there is an increase in grade from the coastal area towards the east in the former area, and southwards from the Orange River in the latter region. In addition, a metamorphic event post-Namaqua, but prior to deposition of the Nama Group occurred along the west coastal region. The exact timing of the latter event in relation to tectonism has not yet been established.

As the aforementioned generalized metamorphic zonation patterns converge on the area of present investigation, the problems here focus on elucidating the relationship between tectonic and metamorphic events occurring prior to, and related to the Pan-African event.

2.7.5 Regional stratigraphic problems

Lithological correlation in the zone separating the Richtersveld from the Bushmanland Subprovince is difficult because:

- most rocks are strongly tectonised, making recognition of protoliths impossible in most cases;
- prograde and retrograde metamorphic events have obliterated earlier mineral assemblages, especially in supracrustal rocks.

The following regional problems require investigation:

(i) If rocks belonging to the Vioolsdrif and Gladkop Suites (Fig. 2.5) can be identified in the present area, this could be used for correlation purposes. Westward continuity of orthogneisses of the Gladkop Suite from the Namaqua geotraverse should be proved, as the Noenoemaasberg and Steinkopf gneisses form only small units in the extreme west of the geotraverse (Fig. 2.5). Theart (1980) mapped the former as sillimanite-bearing nodular gneiss and leucocratic gneiss of the Rietkloof Formation, which shows continuity westwards to the 17°30′ boundary, and can therefore be expected to continue into the present area. The Steinkopf gneiss is mapped as finely laminated gneiss with intercalated amphibolite (Theart, 1980), and is sporadically developed up to the same boundary.

Identification of Vioolsdrif leucogranites in the study area could lead to correlation with those in the Namaqua geotraverse. A leucocratic granite at the southern contact with the Groothoek Formation in the latter area is mapped as a muscovite biotite gneiss (Theart, 1980) showing continuity as a concordant unit to the western boundary of his area (i.e. 17°30′). It is, however, offset by major faults near this boundary, presenting a problem of correlation with lithologies in the present study area;

(ii) The westward extension of the Eyams granite (Fig. 2.5) needs to be proved. Theart's (1980) map suggests that outcrops of this granite can be expected westwards of 17°30′. Joubert (1971), however, groups all leucogranites in this area into his "aplogranite gneiss";

(iii) The stratigraphic relationship of mafic and ultramafic rocks in the present area needs to be established with those in the Namaqua geotraverse and the Wortel Belt of the Pofadder area.

2.7.6 Concluding statement

The problems outlined in this chapter cover a vast spectrum, and in the time allotted for the present survey it became obvious that satisfactory conclusions could not be reached on all of them.

An investigation of literature pertinent to the Namaqua Province has nevertheless been a fruitful exercise as it has highlighted certain geologic problems. The latter arise through controversial interpretations of tectonic and metamorphic relationships between the Richtersveld and Bushmanland Subprovinces.

The specific aims of this study are therefore directed towards:

- detailed geological mapping of an area traversing the western-most part of the marginal zone between the Richtersveld and Bushmanland Subprovinces. The purpose is to document the spatial relationships of lithologies of both subprovinces and to attempt to resolve the problem of how the two subprovinces relate to each other. The present study area was chosen because the Groothoek Thrust zone is well exposed in the Namaqualand/Richtersveld escarpment, and outcrops are good compared to those to the east in the Namaqua geotraverse and farther eastwards in the Geselskap-bank area (Strydom, 1982). In the latter area, in particular, overburden covers key areas critical for interpreting structural relationships along this significant tectonic lineament;

- structural analysis of the chosen area. The purpose of this analysis is to unravel the Early to Mid-Proterozoic and subsequent Late Proterozoic to Early Palaeozoic (Pan-African) structural history of the area;

- determining the metamorphic grade imprinted during the Early to Mid-Proterozoic and Pan-African events in the study area;

- proposing geotectonic models to explain the evolution of the major tectono-thermal events in the region.

3 STRATIGRAPHY AND LITHOLOGY

The investigation required that the spatial distribution and relationship between all rock-types be known for the tectonic history to be deciphered. It was necessary to subdivide the supracrustal rocks into mappable units or formations, in accordance with South African Commission on Stratigraphic Nomenclature principles (SACS, 1980). However, it is not appropriate to give formation status to pre-Gariep high-grade metamorphic rocks in this region because of their disruption by abundant granitic sheets. The term "suite", equivalent to a lithostratigraphic formation in rank is used instead in this study, defined as "related lithologic units characterised by distinctive lithologic, mineralogic, textural or chemical features" (SACS, 1980, p. 650).

Formation names, where used in the literature, have been retained, but others are given after local farms or prominent physiographic features for descriptive purposes specific to this project (see Annexure 1). New names are used where correlation of lithologies is uncertain, and they therefore do not necessarily merit formal status.

Supracrustal rock-types of the Bushmanland Subprovince crop out in the southern strip of this area, namely the Chabiesies suite of the Een Riet Subgroup. Intrusive rocks include mostly granitic types and these form the greater proportion of lithologies in the south. Supracrustals of the Richtersveld Subprovince are found in the northern and central strip and are mostly Windvlakte lavas of the Orange River Group (cf. Fig. 2.5). "Transitional" zone lithologies which occur between rocks typical of the Richtersveld Subprovince and those of Bushmanland Subprovince are here subdivided into the Ratelfontein, Groothoek and Groenrivier suites. Intrusive rocks are predominantly granodiorites and associated Vioolsdrif Suite rocks, as well as mafic and ultramafic rocks.

The relationship of metasedimentary/metavolcanic and intrusive rock-types to one another forms a significant aspect of this chapter, especially in view of the importance of their stratigraphic allocation to either the Bushmanland or Richtersveld Subprovinces. An attempt at correlation with adjacent areas is made to establish a regional geological setting, even though the present investigation does not overlap with other surveys carried out east of 17°30′.

Structural symbols used in this chapter, e.g. $S(N)_2$, are defined in Table 4.1 (Chapter 4).

3.1 EARLY TO MID-PROTEROZOIC SUPRACRUSTAL ROCKS

3.1.1 Chabiesies suite (Bushmanland Subprovince)

Lithologies of this suite crop out in the south-western and south-eastern parts of the study area and form < 10% of the outcrop area (Fig. 3.1). They consist of quartzo-feldspathic and biotite gneisses, with minor metapelitic schists, quartzite and amphibolite. The Chabiesies suite is displaced southwards beyond the study area between (i) the Steenbok and Tierkloof Shears, and (ii) the Kromnek and Stinkfontein Contact Shears (Annexure 1).

3.1.1.1 West of Tierkloof Shear

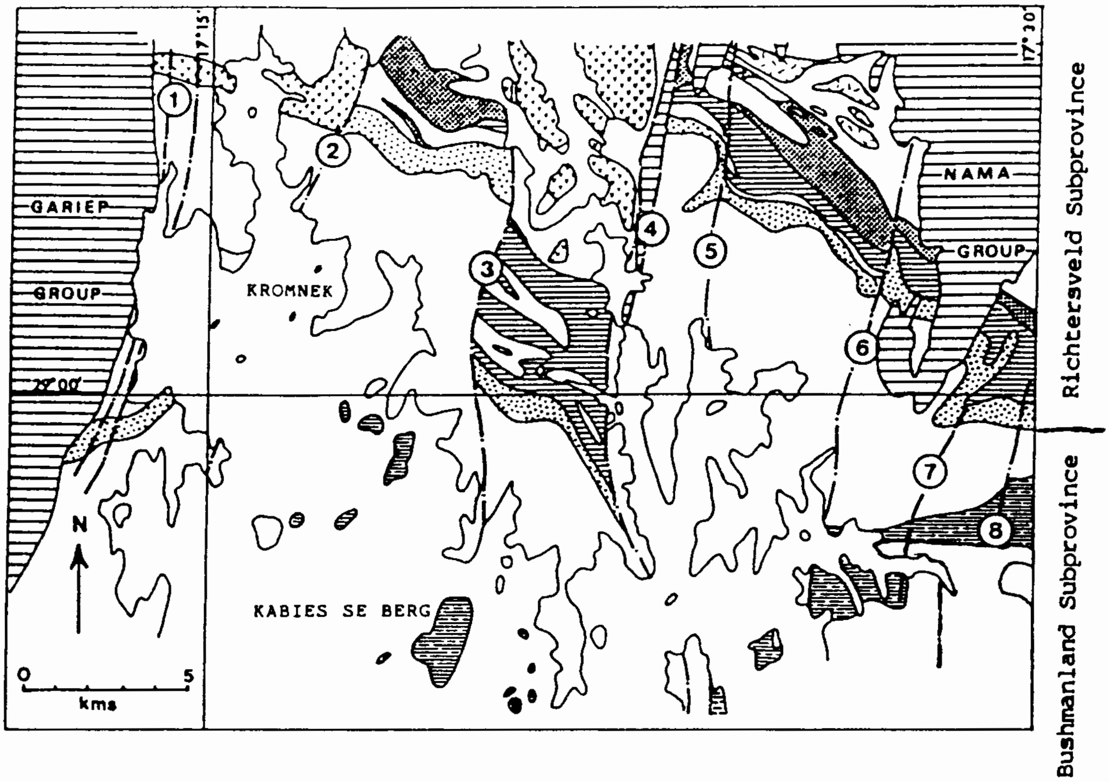
The Chabiesies suite is found in inselberg type outcrops such as Kabies se Berg (Grid reference I12 and J13, Annexure 1). Here, where the succession is at least 380 m thick, it occurs in a northward plunging synform. The rocks are very coarse-grained, and usually banded. North-west of Kabies se Berg rock-types include grey, fine-grained biotite gneisses with intercalated white quartzite bands, occurring as inliers between reddish windblown sand and rock scree. Their strike in this immediate area is mostly north-northeasterly, which is distinctly different from that of other lithological units immediately to the north.

Biotite-garnet-kyanite-staurolite schist

Although the bt-ga-ky schists are a very small volumetric proportion of the sequence, they are a key to understanding the metamorphic and structural events in this area. Their main petrographic features are summarized in Table 3.1. A detailed account of the relationships between minerals is given in Chapter 5.

Quartz commonly forms ribbons averaging 0,5 mm wide (Fig. 3.2 and Plate 1).

Garnet poikiloblasts occur in two predominant forms:
- rounded to elliptical shape, enclosing smaller rounded recrystallized quartz grains, sometimes stubby laths of biotite and muscovite in a typical sieve structure (Spry, 1969);
- oval shape with enclosed minerals as before, but elongated along the foliation and apparently developed during mylonitization of the rock. The enclosing foliation, defined by quartz ribbons, kyanite and phyllosilicates, wraps around the garnet porphyroblasts, suggesting a syn-deformational origin (see Fig. 3.2, and Chapter 5 on metamorphism).



NAMA AND GARIEP GROUPS, Cover rocks		
WINDVLAKTE SUITE, upper unit		RICHTERSVELD SUBPROVINCE
WINDVLAKTE SUITE, lower unit		
GROENRIVIER SUITE	Transitional Zone	
GROOTHOEK SUITE		
RATELFONTEIN SUITE		
CHABIESIES SUITE		BUSHMANLAND SUBPROVINCE

1 Stinkfontein contact shears
2 Kromnek Shear
3 Tierkloof Shear
4 Steenbok Shear
5 Kouefontein Shear
6 Riethoek Shear
7 Chabiesies Shear
8 Witbakensberg Shear

Fig. 3.1 Supracrustal rock-types in the study area.

TABLE 3.1 Petrographic Description of bt-ga-ky-st schist of the Chabiesies suite

Texture: Heteroblastic.

Quartz: Up to 50% by volume, individual grains have notably inequant shape, strain shadows usually present, some degree of suturing between grains.

Feldspar: Oligoclase, present in varying proportions, seldom exceeding 25% by volume, anhedral grains, variable size, usually smaller than quartz grains.

Mica: Brown biotite and muscovite, broad laths about 1 mm long, usually parallel to foliation but quite often randomly oriented

Garnet: Conspicuous, especially in certain layers within more pelitic rocks; habit ranges between mimetic growth along foliation, where it may enclose blades of kyanite and rounded smaller grains of quartz, to large poikiloblasts up to 6 mm size.

Kyanite: Narrow blades and thin individual needles usually no longer than 0,5 mm, occurring predominantly parallel to the foliation.

Staurolite: Abundant, usually confined to particular layers within the rock, crystals vary in grain-size depending on orientation, smaller euhedral grains average 1 mm, grains developed along foliation can be up to 5 mm long.

A notable characteristic of staurolite is its post-tectonic development, noted earlier by Rogers (1915, p. 75) and Joubert (1971, p. 32). Grains show the following habits:
- poikiloblastic texture, where small rounded quartz grains are enclosed within the larger, often euhedral crystal. This habit is not as well developed as in staurolites found farther north-west, in the stratigraphically higher Ratelfontein suite;
- enclosure of an earlier foliation, Si. The latter is continuous with the external foliation, Se (see Fig. 3.2 and Plate 1).

Both features indicate post-tectonic staurolite growth and are important in relating tectogenesis to metamorphism (see Chapter 5).

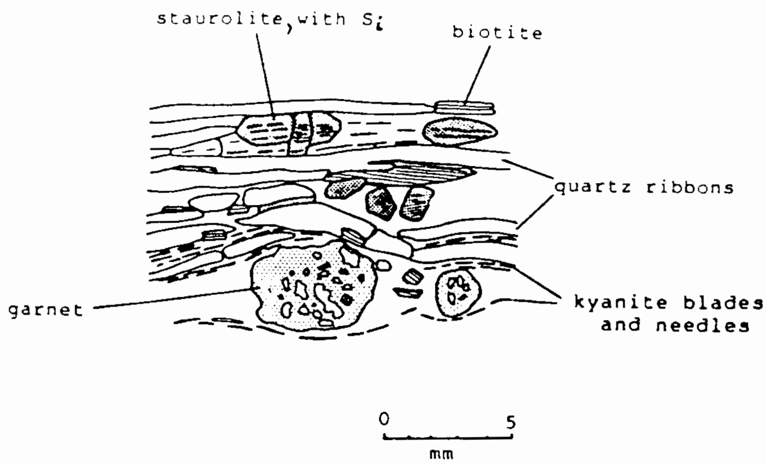


Fig. 3.2 Mylonite texture in metapelite, Chabiesies suite. Note how garnet porphyroblasts "bow-out" the main foliation, and post-tectonic staurolite retains Si. Drawn from a photomicrograph.

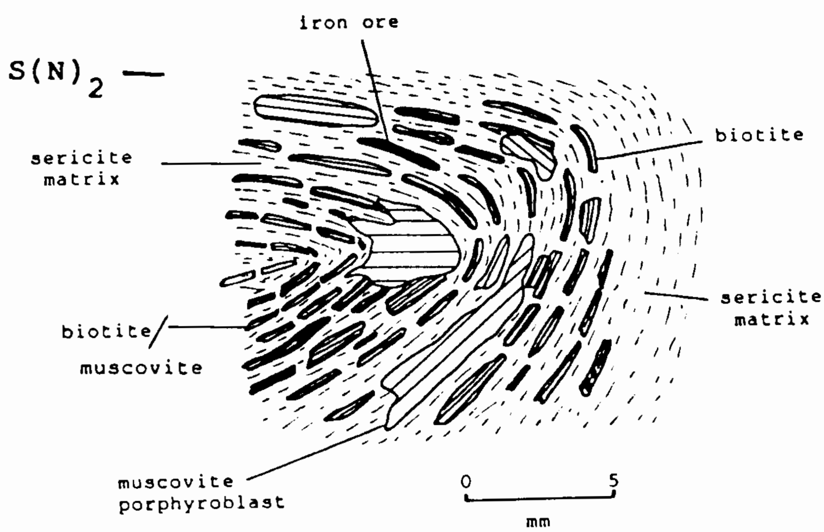


Fig. 3.3 Tight fold in a biotite schist, Chabiesies suite. Note muscovite porphyroblasts in hinge zone of fold, and a mimetic development in limbs of fold. Note also advanced replacement of all minerals by sericite. Drawn from a photomicrograph.

Kyanite is generally associated with staurolite, having developed at the same time. It forms small subhedral laths seldom more than 1 mm, but occasionally up to 4 mm long. It has a random orientation but, like staurolite, is confined to certain layers, seldom transecting the dominant quartz ribbon texture of these metapelites. Kyanite does not develop as abundantly as on the farm Steenbok 50 km to the south, or reach the large crystal size noted on Nakanas, west of Steenbok, by Rogers (1915) and Joubert (1971). Characteristically these schists possess a penetrative cleavage, usually in an anastomosing pattern. Small stubby muscovite crystals, averaging 0,05 mm long, define this foliation. In some thin sections however, muscovite shows a lath-to needle-like, cleavage-parallel habit. Along with biotite, it is the dominant mineral of these pelitic rocks. This cleavage is the same in pelitic rocks east of the Steenbok Shear, but there it is outlined by very fine-grained white mica (sericite).

Quartzo-feldspathic gneiss

This is a beige, leucocratic, coarse-grained rock, with the foliation enhanced by the presence of biotite.

Quartzite

Approximately north-striking, whitish, thin beds of recrystallized quartzite are present at several levels in the sequence. Towards the base they are less continuous than higher up, where they form positive weathering features across Kabies se Berg.

Biotite-garnet schist, biotite schists and porphyroblastic gneisses

Towards the base of the sequence, bt-ga schists and bt schists form thin discontinuous beds within banded and minor porphyroblastic gneisses.

3.1.1.2 East of Steenbok Shear

South-east of Steenbok Shear the Chabiesies suite occurs as segmented units in orthogneisses, making it difficult to assess its total thickness. It is comprised of gneissic, schistose and quartzitic rock-types.

Quartz-feldspar gneiss

This rock-type is found near the top of the succession, where occasional biotite schist bands are present. A leucocratic, foliated variety crops out east of the old Chabiesies homestead. The main petrographic features of this rock are summarized in Table 3.2.

TABLE 3.2 Petrographic Description of quartz-feldspar gneiss of the Chabiesies suite.

Texture: Equigranular; approximately equal proportions of plagioclase (oligoclase) and quartz, minor amounts of microcline, muscovite, biotite, chlorite, epidote and sericite*.

Oligoclase: grain-size generally < 0,7 mm, mostly in advanced stage of saussuritization.

Microcline: small anhedral unaltered grains, < 0,3 mm size.

Quartz: anhedral, grain-size < 0,8 mm, undulose extinction common, sutured grain boundaries only occasional.

Biotite: characteristic colour deep brown, occasionally olive- green, forms skeletal intergrowth with quartz.

Muscovite: small laths, edges replaced by sericite.

Sericite: present in entire fabric parallel to the foliation, can constitute up to 30% by volume.

Chlorite: anomalous interference colours common, occurs as discrete grains and as replacement of biotite.

Accessories: small anhedral epidote and irregularly shaped iron-ore grains.

* The term "**sericite**" is used throughout this text to describe a fine-grained flecky occurrence of white mica (muscovite) which texturally differs from muscovite laths and porphyroblasts.

In places sericite occurs in elongate needle-like form within quartz grains. This texture is interpreted as a replacement of sillimanite which originally occurred in needle-like habit in the quartz-feldspar gneiss. Virtually every mineral shows varying degrees of replacement by sericite and/or chlorite. The larger quartz grains are broken up into smaller grains along cracks, while phyllosilicates show advanced replacement textures.

Biotite-muscovite schist

Bt-mu schist beds usually about 1 m thick crop out, for example, just east of the old Chabiesies homestead. Their colour is dark green, and they usually display tight to isoclinal folds.

Petrographic features are summarized in Table 3.3.

TABLE 3.3 Petrographic Description of bt-mu schist of the Chabiesies suite

Texture: Inequigranular; chiefly muscovite, biotite, quartz and sericite with a small amount of plagioclase, iron ore and chlorite.

Muscovite: Two generations, first has form of thin laths representing the earliest recognizable foliation folded into tight folds (Fig. 3.3), second is large muscovite porphyroblasts transgressing tight fold structures and growing mimetically along pre-existing foliation on fold limbs (Fig. 3.3). Porphyroblast habit sometimes poikiloblastic, with undulose extinction, up to 4 mm long.

Biotite: Brown, lath-shaped grains, outlines earlier foliation, marginally replaces muscovite.

Chlorite: Replaces biotite along the margins.

Sericite: Fine-grained flecky appearance, replaces phyllosilicates along foliation. Accounts for about one third of the volumetric proportion of minerals present.

Accessories: Minor epidote as rounded anhedral grains; iron-ore notably replaced by sericite, strung out along the foliation.

Muscovite and biotite laths outline $S(N)_2$ which is folded into tight folds (Fig. 3.3; and see Table 4.1). The advanced stage of replacement of pre-existing minerals, especially the phyllosilicates, by sericite and chlorite is notable. In some cases about one third of the rock is made up of the latter two minerals. Sericite particularly forms approximately 2 mm thick anastomosing layers. The latter mineral formed late in the history of this rock (D(N)3 tectonic episode, (Chapter 4)), along a prominent cleavage throughout the rock.

Quartzite

Two prominent quartzite units constitute the high ground at Beacon 31 in the south-east, and a thinner unit is located just over 2 km south-west of the old Chabiesies homestead. These quartzites are whitish to light grey in colour, recrystallized, and devoid of primary structures. They form continuous conformable units within biotite gneisses and schists.

Amphibolite

Amphibolite is a very minor component of the sequence. It usually occurs as thin discontinuous bands in the biotite gneisses. It is usually dark green to black in colour, relatively coarse-grained, composed chiefly of hornblende, plagioclase, epidote, minor apatite and quartz. The amphibolite layering is particularly noticeable under the microscope. Leucocratic layers are composed of plagioclase grains (up to 2 mm), separated from the darker bands containing a mixture of equigranular, granoblastic-polygonal blue-green hornblende and plagioclase grains. Saussuritization of feldspars is obvious in these rocks. Epidote is abundant, some at least being derived from the break-down of hornblende. A minor amount of quartz and apatite is also present.

Biotite-garnet-chloritoid schist

This occurs in very thin beds composed mainly of biotite with occasional quartz, garnet and muscovite. Nearly all minerals have been replaced to some degree by sericite and chlorite, leaving isolated palimpsests (see Fig. 3.4).

Microscopically, the rocks show broken quartz grains replaced by sericite. Biotite laths are randomly oriented, some completely replaced by chlorite. Muscovite porphyroblasts are present, largely replaced by fine-grained sericite.

Some remnant garnet porphyroblasts are present. In places, broken and replaced by chlorite and sericite, these are strung out along the foliation, along with irregular shaped grains of iron ore.

Sericite makes up about 60% of the rock and occurs along a well developed anastomosing foliation. Within one particular horizon in the south-east of the area the prominent sericitic foliation has been deformed into isoclinal folds (grid reference S 12, Annexure 1). Chloritoid has grown across these folds as stubby lath-shaped crystals, but these in turn have been slightly buckled (Fig. 4.52, Chapter 4).

In places sericite has a similar habit to that occurring in quartz feldspar gneisses of this suite. This texture is interpreted as a replacement texture, sillimanite needles having been completely

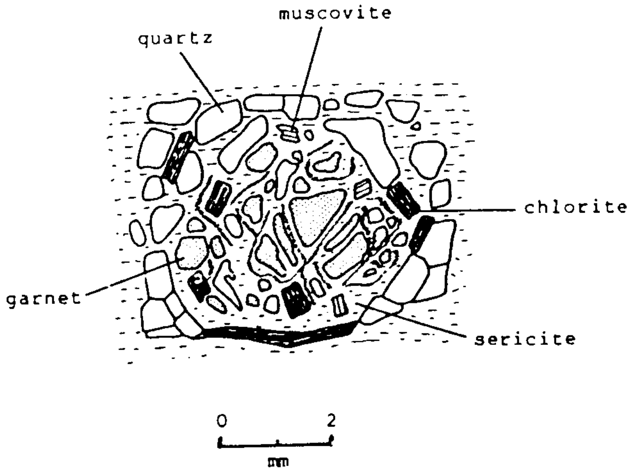


Fig. 3.4 Garnet and biotite replaced by sericite and chlorite in biotite-garnet-chloritoid schist, Chabiesies suite. Drawn from a photomicrograph.

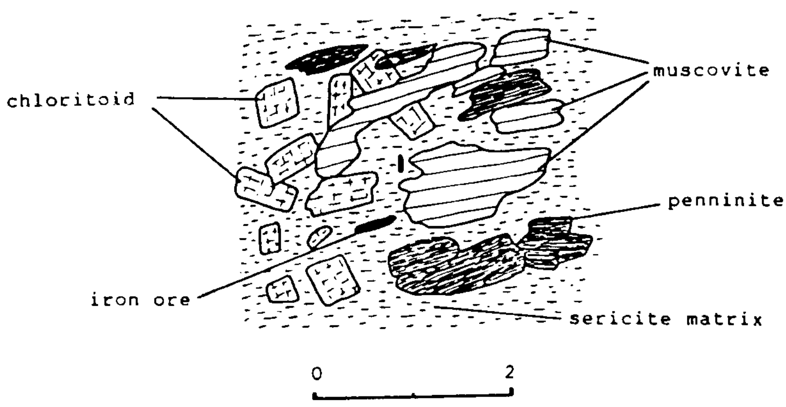


Fig. 3.5 Sericite replaces large muscovite porphyroblasts and broad bands of sillimanite in a metapelite, Chabiesies suite. Chloritoid and penninite show post-tectonic growth. Drawn from a photomicrograph.

replaced by sericite. Similarly, prominant wispy forms of sericite occurring in layers within this rock-type are interpreted as replacement of sillimanite. This interpretation is valid because sillimanite occurring in the latter form is common in metapelites of the Bushmanland Subprovince. This texture has been recorded at Geselskapbank, east of the study area by Grütter (1986), and in supracrustals of the Aggeneys area (P. Smith, personal communication, 1987). In addition, garnet-biotite geothermometry determinations on rocks in the eastern part of the study area confirm that these rocks were elevated to temperatures appropriate for the development of sillimanite in metapelites (Chapter 5).

Xenolithic fragments within the Kouefontein granite gneiss

About 3 km north of Chabiesies large xenolithic blocks of pelitic, psammitic and amphibolitic rock-types occur within the Kouefontein granite gneiss. The fabric in the xenoliths is continuous with that in the host granite gneiss. The xenoliths are made up mostly of schistose rock-types.

Muscovite-quartz-chlorite schist

Quartz makes up about one third of the rock, and displays deformation features such as mortar texture and quartz ribbons up to 0,5 mm wide. Muscovite is present in two forms: (a) smallish broad laths in skeletal intergrowth with quartz; (b) large porphyroblasts of a later generation. Chlorite (penninite) has replaced the first generation muscovite sometimes markedly, whilst apparently not affecting the larger porphyroblasts. Sericite conspicuously replaces the phyllosilicates throughout the rock. Accessory minerals include epidote, iron ore and apatite.

Biotite-muscovite-sericite schist

Brown biotite and muscovite are the major constituents, developed as particularly large porphyroblasts, mostly oriented parallel to the foliation and showing undulose extinction. Minor amounts of quartz and iron ore are present, the latter both in euhedral forms and as anhedral replacement of biotite and muscovite. Up to 60% by volume is composed of sericite which defines the strongly developed cleavage throughout the rock.

Muscovite-sericite-chlorite-chloritoid schist

This is composed chiefly of chlorite, muscovite, abundant fine-grained sericite and chloritoid. Quartz is notably absent. Large muscovite porphyroblasts have grown across the felted mass of sericite, in places preserving the needle-like and wispy shape of what appear to be former sillimanite grains. Chloritoid similarly has grown across the sericite groundmass in clusters of randomly oriented euhedral crystals averaging 1 mm in size (Fig. 3.5).

Multiple twinning and strongly developed green and blue pleochroism characterise this mineral in thin section. Euhedral chlorite (penninite) crystals have grown across the muscovite porphyroblasts and within the matrix, imparting an overall green colour to the rock.

Biotite-quartz-feldspar gneiss

Similar to the gneisses farther south, this coarse-grained rock is composed predominantly of quartz, plagioclase and microcline, with minor biotite (in places completely replaced by chlorite) and epidote.

Hornblende gneiss

In field appearance this rock resembles that described farther south on Chabiesies. It forms relatively thin units in this eastern outcrop area never more than a few metres thick. It is a dark green to black, medium to coarse-grained rock, often showing isoclinal and open folding.

Inequigranular throughout, it is composed mainly of amphibole with lesser amounts of quartz, muscovite, epidote, apatite and iron-ore. It has a faintly banded appearance due to alternating amphibole-rich and leucocratic mineral layers. Blue-green hornblende, up 80% of this rock, is present as euhedral grains, generally less than 1 mm in size. Muscovite forms irregular, generally lath-shaped subhedral grains, less than 0,2 mm in size. Small anhedral grains of quartz, epidote, iron ore and apatite are interspersed among the amphiboles.

Mineral assemblages of diagnostic metapelites and metabasites of the Chabiesies suite are summarized in Table 3.4.

3.1.1.3 Correlation

The Chabiesies suite shows a divergence in lithological and structural trend when compared to the rock-types to their immediate north. This pattern continues for at least 25 km east of the area (cf. Fig. 2.5).

Chabiesies suite rocks can be correlated with part of the Rietkloof Formation east of 17°30′ (Theart, 1980; see Fig. 3.6). Theart includes rock-types such as sillimanite-bearing nodular gneiss and leucocratic gneiss (Map code sg2), finely laminated gneiss with intercalated amphibolite (lg), metaquartzites (mq) muscovite-gneiss (mg) and garnet-biotite schists (sm) in the Rietkloof Formation, part of the Een Riet Subgroup. Blignault et al. (1983), however, differentiate the supracrustals into the

TABLE 3.4 Mineral assemblages of pelitic and amphibolitic rocks of the Chabiesies suite.

Sample No.	quartz	plagioclase	biotite	garnet	kyanite	staurolite	muscovite	sericite	chlorite	chloritoid	apatite	hornblende	epidote	iron ore	Locality re: Steenbok Shear	Grid ref. Annexure 1
PB476*	x	x	x	x	x	x	x								west	I 12
PB 79*							x	x	x	x					east	Q 9
PB 90*	x	x	x				x	x	x					x	east	S 11
PB462*	x		x	x			x	x	x	x					east	T 12
PB409⊕	x						x				x	x	x	x	east	T 9
PB624+	x						x					x	x		east	S 11

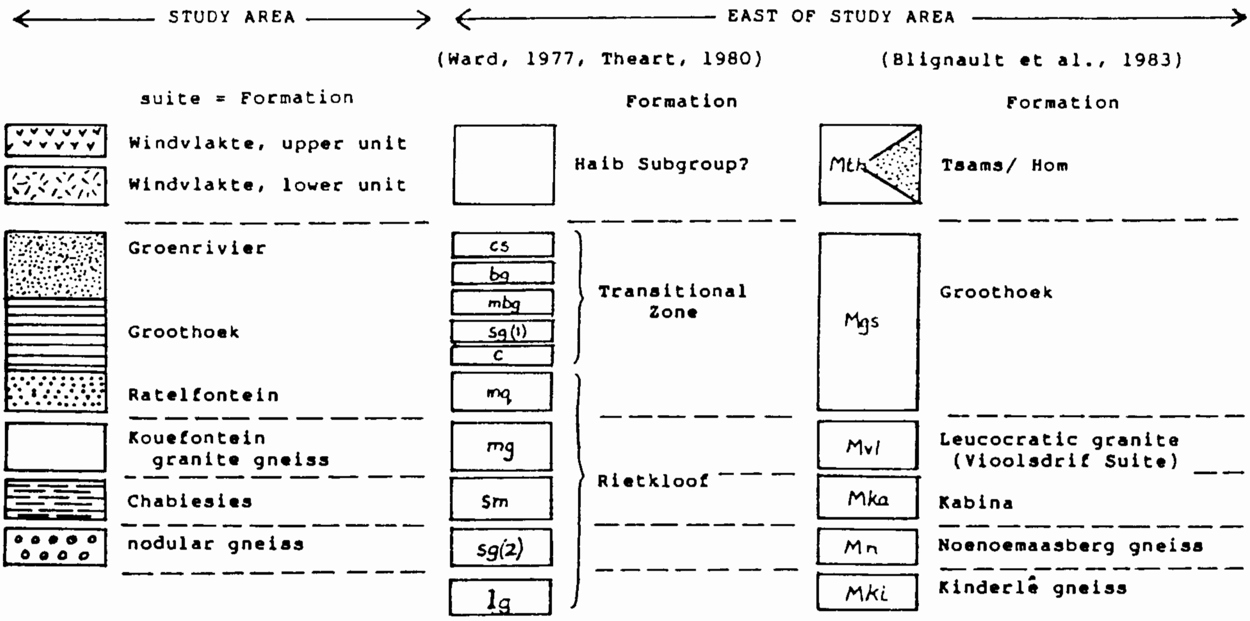
* = metapelites

\+ = amphibolite

⊕ = hornblende gneiss

formations Tsams/Hom, Groothoek and Kabina (cf. Fig. 2.5). They name the muscovite gneiss (mg) of Theart a leucogranite (Vioolsdrif Suite) and the nodular gneiss (Sg2) as the Noenoemaasberg gneiss. Theart's finely laminated gneiss (lg) is a possible correlate of the Steinkopf gneiss.

In the study area the Chabiesies suite is correlated with the Kabina Formation of Blignault et al. (1983), based on similar lithologies.



Map codes, Theart (1980)

cs	nodular cordierite schist
bg	fine-grained biotite gneiss
mbg	muscovite-biotite gneiss
sg(1)	schistose biotite-muscovite gneiss and schist
c	carbonate band
mq	metaquartzite and quartzitic schist
mg	muscovite gneiss
sm	garnet biotite schist, sillimanite-bearing & graphite-bearing schist
sg(2)	sillimanite-bearing nodular gneiss and leucocratic gneiss
lg	finely laminated gneiss with intercalated amphibolite

Map codes, Blignault et al. (1983)

Mth	undifferentiated volcanics, nodular cordierite schist
Mgs	mica quartz schist with muscovite-cordierite-sillimanite layers, calc-silicate layers, amphibolite, metaquartzite, cordierite-anthophyllite rock
Mvl	leucocratic granite
Mka	metaquartzites, biotite-muscovite-sillimanite schists, magnetite quartzite, amphibolite, cordierite-anthophyllite rock
Mn	reddish brown weathering leucocratic gneissic granite
Mki	leucocratic, reddish brown weathering quartzo-feldspathic gneiss & sillimanite nodules

Fig. 3.6 Correlation of supracrustal lithologies west of 17°30' with those to the east. Map codes taken from Theart (1980) and Blignault et al. (1983).

3.1.2 Ratelfontein suite (Richtersveld Subprovince)

This suite, some 500 m thick, corresponds to the "Quartzite Zone" of the "Pink Gneiss Unit" (Ritter 1980, p. 70), described as "muscovite biotite schist, quartzite embedded in pink gneiss and muscovite schist". De Villiers and Söhnge (1959, p. 33) describe these quartzites, which strike predominantly north-west, and note that this consistent attitude is disrupted at intervals where the strike changes to north/south, or the quartzites form small outliers.

During the present investigation the quartzite/ schist association was found to persist across the entire region, disrupted at intervals by north-northeasterly trending (Pan African) shears. The quartzite/schist horizon marks the northern-most outcrop of quartzite within the supracrustals of this area, and is a useful marker for mapping, and in solving stratigraphic and structural problems. The north-westerly regional trend of these rocks contrasts with the variable trend of lithologies noted in the Chabiesies suite to the south.

3.1.2.1 Lithologies and field description

Biotite-muscovite-quartz-feldspar gneiss

This occurs at or near the base of the unit, and intercalated with quartzite slightly higher in the sequence. It may in fact be a deformed contact zone of the underlying Kouefontein granite gneiss.

This coarse-grained rock is composed mainly of quartz and plagioclase with lesser microcline, muscovite, biotite, garnet and sericite. Muscovite porphyroblasts show an advanced stage of replacement by sericite. Sutured boundaries are present between large quartz grains.

Biotite gneiss

A pale grey, fine-grained rock occurring at the base and slightly higher in the sequence, it forms thin discontinuous horizons and was observed only east of the Kouefontein Shear.

Quartzite

Predominantly whitish to pale grey with a glassy recrystallized field appearance, they are composed of quartz with very minor muscovite. Slightly higher, small boudinaged units of a dark grey to blackish quartzite are present east of Kouefontein Shear. The colour change is due to the presence of magnetite. Magnetite quartzites, however, form a very insignificant proportion of this quartzite unit.

Although these northern-most outcropping quartzites are a prominent marker unit, individual mesoscale characteristics are laterally and vertically diverse. Firstly, the quartzites are seldom continuous laterally, but are shaped like segmented "pips" or boudins (Fig. 3.7). Boudin length averages a few tens of metres, thickness between 0,5 and 2 m. Along strike their length may reach 100 m, but this is exceptional. The boudinage is also expressed vertically, occurring en echelon in section where the quartzites are intercalated with gneisses and schists.

One outcrop of quartzite showing tectonic banding was found near the Riethoek Shear (grid reference Q5, Annexure 1). The banding occurs in a greenish quartzite associated with cordierite-anthophyllite and garnet-staurolite metapelite beds near the base of the succession (see Plate 2). In thin section the banding is seen to be due to grain-size refinement, narrow bands of granoblastic-polygonal quartz (grain-size about 0,25 mm) alternating with broad bands (about 3 mm) of coarser polygonal inequant grains (up to 1 mm).

Within this granoblastic-polygonal texture are small stubby brown biotite grains. Chlorite is present between the smaller quartz grains, locally giving the rock a "web texture". A very minor proportion of small rounded epidote grains is also present.

West of Kromnek Shear the Ratelfontein quartzites which crop out at the Penssleep beacon (No.70) trend westerly to south-westerly, an anticlockwise swing from their normal orientation farther east. West of the Penssleep beacon north-striking quartzites of the Stinkfontein Formation overlie them unconformably. The unconformity is clear (Annexure 1) even though the quartzites in this area are of similar field appearance, making it difficult to distinguish between them (Joubert, 1981).

Cordierite-anthophyllite-sillimanite schist

This was observed in one locality, in association with the banded greenish quartzite described above, near the Riethoek Shear (i.e. in the eastern part of the study area). Thin bands within the quartzite are made up of altered cordierite, anthophyllite and biotite. Sillimanite remains as large fragmented porphyroblasts, only partially replaced by sericite. Sericite has completely replaced anhedral grains of cordierite. Anthophyllite has retained its wispy radiating habit, but has been replaced by chlorite. It can occasionally be seen to enclose clusters of quartz, individual sillimanite, and euhedral apatite and garnet grains. A very minor proportion of epidote is also present.

The extensive replacement of most minerals by sericite gives the rock an overall "web texture", especially where the larger prisms of sillimanite and quartz grains are present.

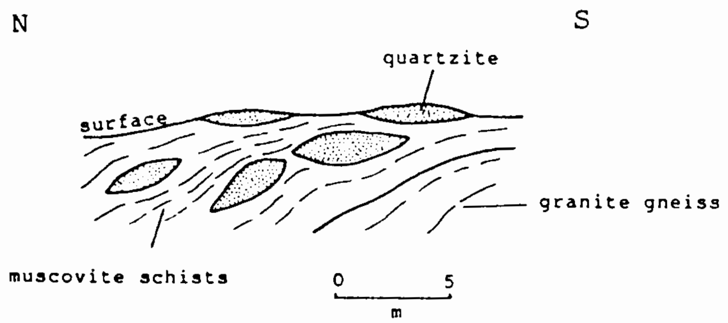


Fig. 3.7 Quartzite boudins in metapelitic rocks, Ratelfontein suite. Drawn from a photograph.

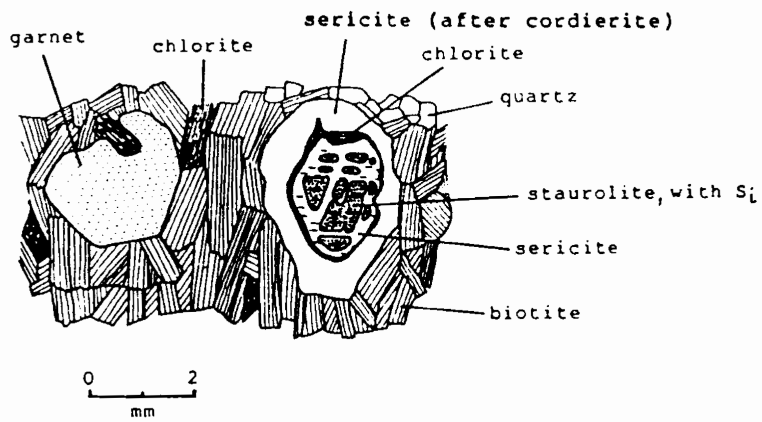


Fig. 3.8 Replacement of cordierite and staurolite by sericite and chlorite in a metapelite, Ratelfontein suite. Drawn from a photomicrograph.

Biotite-garnet-staurolite schist

This was located only east of the Steenbok Shear, as a 20 cm thick horizon occurring interbedded with the greenish quartzite described above (Plate 2). In outcrop it has a dark green to blackish colour with prominent garnet dodecahedra protruding from a groundmass of broad blades of biotite.

A petrographic description is given in Table 3.5.

TABLE 3.5 Petrographic Description of bt-ga-st schist of the Ratelfontein suite

Texture: Coarse-grained, inequigranular.

Garnet: Prominent euhedral crystals up to 10 mm in diameter. Criss-crossed with numerous cracks- some arranged radially around small zircon inclusions. Sericite and chlorite present within the cracks.

Staurolite: Remnant grains, largely replaced by sericite and chlorite.

Cordierite: Hexagonal grains surround staurolite porphyroblasts, but occurs completely pinitized.

Biotite: Forms broad brown blades, up to 5 mm long, constitutes 50% of the rock. Some have pronounced pleochroic haloes. Subhedral grains strongly aligned along $S(N)_2$, but in places give an overall decussate texture to the rock. Most show undulose extinction and kinking.

Epidote: A minor amount present as small anhedral grains, usually less than 0,07 mm size.

Quartz: Forms a very minor proportion of the rock, polygonal-shaped grains usually measuring about 0,5 mm. They occur in single bands of this width, but also in clusters where sutured boundaries are present between grains showing strain shadows.

Chlorite: Occurs as discrete anhedral grains, also replacing biotite, garnet and staurolite.

Sericite: Forms up to 20% of the rock, usually as a felted mass of small laths which replace biotite, garnet and staurolite.

Note that where chlorite forms a distinct rim to the former edge of staurolite grains it shows that staurolite porphyroblast size ranged up to 4 mm (Fig. 3.8). Some individual porphyroblasts contain inclusions trails of opaque material (either graphitic or ore) - an expression of an earlier S_1 feature. Staurolite is always enclosed by either cordierite or garnet, and has therefore survived as an armoured relic, during the Namaqua metamorphic event (Chapter 5).

A chemical analysis reveals a high abundance of major oxides FeO and MgO, as expected from the mineralogical composition. The rock is, however, low in K_2O, Na_2O, CaO and MnO. Among the trace elements only Zn and possibly Zr abundances are slightly anomalous (see Table 3.6, Sample PB 510).

Muscovite schist

Overlying and intercalated with the quartzites, these are beige, medium-grained schists with prominent muscovite porphyroblasts defining the schistosity.

The rocks are composed of varying proportions of essentially muscovite and quartz, with lesser plagioclase, chlorite and sericite. Muscovite, up to 45% by volume, has a definite preferred orientation parallel to the regional foliation, occurring as broad lath-shaped blades interspersed with inequant quartz grains. Most of the muscovite shows variable undulose extinction, due to kinking and straining. Occasional large porphyroblasts up to 5 mm long overgrow the strongly foliated fabric.

Chlorite is sometimes present as individual wispy fibrous crystals averaging 6 mm in size, as well as replacing phyllosilicates. Sericite is usually present along the foliation in varying proportions, where it replaces largely muscovite and chlorite.

Muscovite-quartz schist

In these rocks quartz is present in greater proportions than in the previous type. Most grains are inequant, their long axes aligned parallel to the $S(N)_2$ foliation. A petrographic description is given in Table 3.7.

Quartz-muscovite-biotite-staurolite-garnet schist

These rocks overlie prominent quartzites at the Penssleep beacon (No.70), near the contact with the Stinkfontein Formation, in the extreme west of the area. They are the lateral equivalents of the semi-pelitic and pelitic rocks described above, but differ slightly in their mineralogy. They are essentially qtz-mu schists, but have a conspicuous nobbly texture caused by protruberance of garnet and staurolite porphyroblasts above the weathering surface.

TABLE 3.6: Geochemical analyses of metapelitic rocks. Major element oxides expressed in percentages; trace elements in ppm.

	PB-499	PB-500	PB-510	#1		#2
SiO_2	39.87	68.30	47.00	65.40	(6.03)+	37.82
TiO_2	0.71	0.70	1.47	0.76	(0.17)	0.22
Al_2O_3	27.71	15.59	14.33	17.91	(3.52)	14.72
FeO *	4.88	5.43	19.78	5.33	(2.30)	15.37
MnO	0.09	0.06	0.80	0.10	(0.14)	0.37
MgO	14.54	1.72	9.77	1.21	(0,45)	14.36
CaO	0.04	0.09	1.41	0.18	(0.19)	1.81
Na_2O	0.34	0.72	0.28	0.18	(0.16)	1.10
K_2O	5.10	4.79	1.51	3.91	(1.44)	1.10
P_2O_5	0.05	0.07	0.17	0.05	(0.03)	-
H_2O+	7.62	2.86	3.15	-		7.10
H_2O-	0.23	0.18	0.23	-		6.44
TOTAL	101.16	100.51	99.89			100.41
Rb	189	126	314	218	(87)	
Ba	663	891	452	728	(651)	
Sr	13	34	25	27	(16)	
Th	15	19	29	17	(6)	
U	6	<4	6	-		
Zr	403	235	445	267	(77)	
Nb	11	9	25	16	(3)	
Mo	<2	<2	3	-		
Cr	85	85	81	92	(29)	
V	104	81	111	93	(48)	
Sc	28	13	20	16	(4)	
Ni	45	38	28	23	(15)	
Co	22	17	35	13	(8)	
Pb	9	9	9	21	(19)	
Zn	65	34	290	127	(160)	
Cu	<1.5	1.5	6.3	-		
Y	36	36	43	27	(8)	
La	19	49	47	26	(10)	
Ce	39	98	104	55	(17)	
Nd	21	50	51	27	(11)	

Sample No.	Rock-type, and origin	Grid ref. Annex. 1
PB499	phlogopite-corundum schist, Groenrivier suite	O2
PB500	cordierite schist, Groenrivier suite	O2
PB510	biotite schist, Ratelfontein suite	Q4
#1	Average metapelite from 35 samples, northern Namaqualand (Moore, 1986)	
#2	Chlorite-montmorillonite alteration product of tuff and tuffaceous sediment, Japan (Kimbara et al., 1971).	

(Analyst Dr. J.M. Moore, Dept. of Geology, University of Cape Town)

* Total Fe as FeO + Standard deviation

TABLE 3.7 Petrographic Description of mu-qtz schist, Ratelfontein suite

Texture: Inequigranular.

Quartz: Irregular oval-shaped large grains often display strong undulose extinction, separated by sutured boundaries. These grains are surrounded and interspersed with smaller recrystallized polygonal quartz sub-grains averaging 0,05 mm in size. Large grains sometimes stretched into ribbons, separated by a felted mass of sericite matrix up to 1,5 mm wide. Ribbons often composed of smaller recrystallized quartz grains, which join up with the larger inequant quartz grains (Fig. 3.9).

Muscovite: Conspicuous as large subhedral to irregularly shaped porphyroblasts oriented parallel to the foliation.

Accessories:
- Stilpnomelane, as subhedral grains aligned parallel to the foliation. Has characteristic golden-yellow to reddish-brown pleochroism;
- Magnetite, forms euhedral as well as irregularly shaped grains along the foliation;
- Chlorite, appears as discrete grains in the matrix, and as aggregates where it has replaced iron-ore. It also occurs interleaved with stilpnomelane;
- Sericite, prominent as a late stage mineral replacing phyllosilicates, and enhancing tectonic fabric.

A prominant foliation ($S(P)_2$) paralleling the trend of Pan-African shears is defined by small muscovite laths which transect the granoblastic-polygonal quartz and muscovite groundmass (Fig.3.10). Large euhedral tabular chlorite porphyroblasts (after chloritoid) have developed across this fabric.

A later muscovite cleavage ($S(P)_4$) has developed at an angle of about 40° to $S(P)_2$, locally transecting this foliation, causing a buckling and kinking of the chlorite porphyroblasts.

A petrographic description is given in Table 3.8.

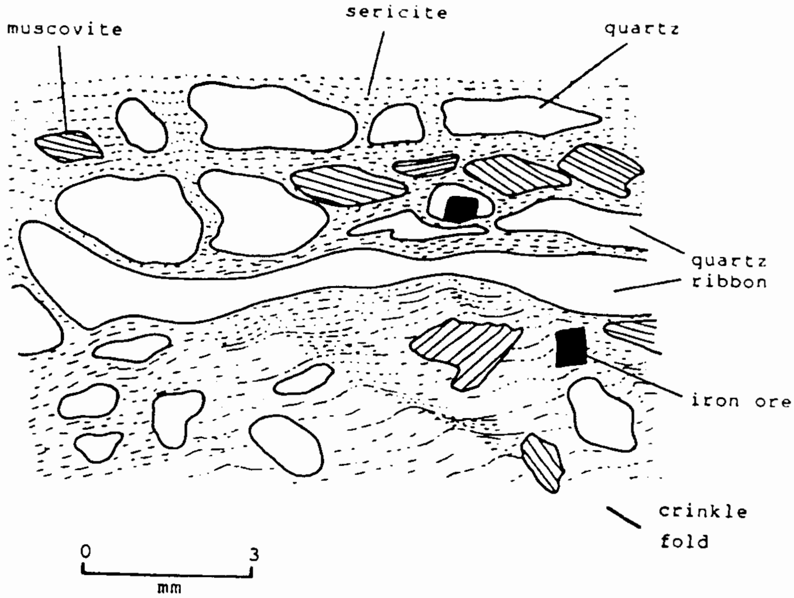


Fig. 3.9 Mylonitic textures in semi-pelitic rock, Ratelfontein suite. Drawn from a photomicrograph.

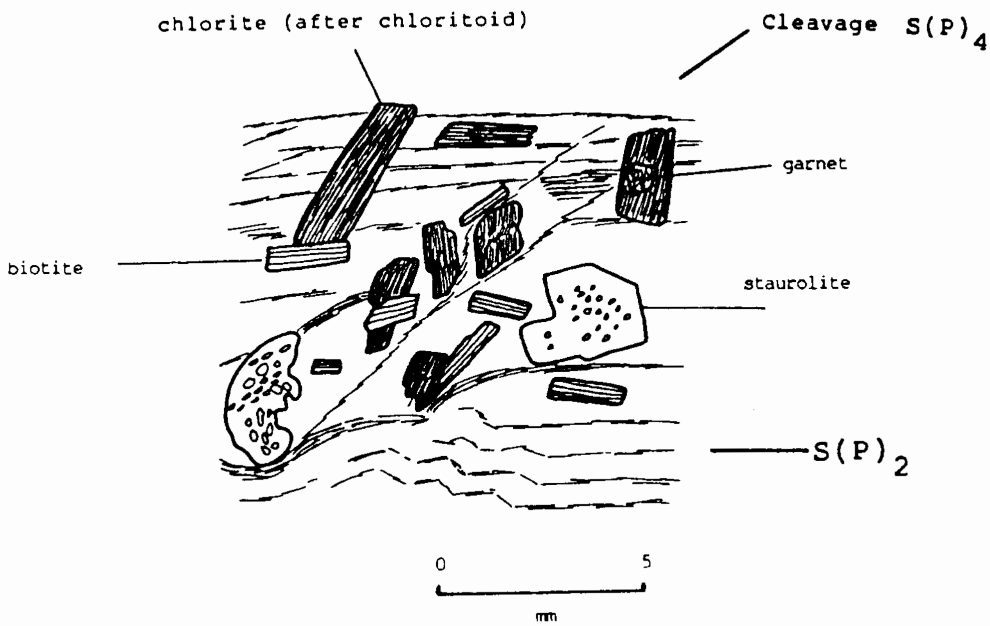


Fig. 3.10 Metapelite from the western extremity of Ratelfontein suite showing porphyroblasts of chlorite, biotite, garnet and staurolite which developed later than $S(P)_2$. Chlorite has completely replaced chloritoid, and in places shows kinking by the $S(P)_4$ cleavage. Sketched from a thin section.

TABLE 3.8 Petrographic Description of qtz-mu-bt-ga-st schist, Ratelfontein suite

Texture: Inequigranular; large, randomly oriented chlorite, biotite, staurolite and garnet porphyroblasts developed within a granoblastic-polygonal groundmass of quartz and small muscovite laths (Fig. 3.10).

Chloritoid: Locally up to 5 mm long, replaced by chlorite; retains certain original chloritoid characteristics e.g. twinning and muscovite inclusions within their inner zones.

Biotite: Brown, occur as porphyroblasts which transect original chlorite porphyroblasts.

Staurolite: Forms prominent poikiloblasts up to 5 mm across. Cores of some larger grains show abundant quartz inclusions surrounded by larger zones with a lesser number of inclusions (cf. Joubert, 1971). Most poikiloblasts are post-tectonic, having grown across the $S(P)_2$ foliation, but some show syn-deformational textures within their quartz-rich cores.

Garnet; Porphyroblasts, corroded around their edges, although some subhedral grains are present.

Chlorite: Occurs as discrete crystals, as portion of the groundmass and in replacement of biotite and garnet.

Mineral assemblages of pelitic and semi-pelitic rocks of the Ratelfontein suite are summarized in Table 3.9.

3.1.2.2 Correlation

The Ratelfontein quartzite/schist association is persistent across the entire study area. As in the case with the Chabiesies suite the western-most outcrops of the Ratelfontein suite show mineral assemblages and textures which differ from those in it's eastern-most part. Garnet-staurolite-chloritoid assemblages west of Kromnek Shear are interpreted as medium-pressure - medium temperature assemblages formed during the Pan-African event. Assemblages east of Steenbok Shear are interpreted as having reached middle amphibolite facies during the Namaqua event, at 1200 Ma (Chapter 5).

TABLE 3.9 Mineral assemblages of pelitic and semi-pelitic rocks of the Ratelfontein suite.

Sample No.	quartz	muscovite	biotite	staurolite	garnet	chlorite	sericite	epidote	stilpnomelane	iron ore	sillimanite	apatite	cordierite #	anthophyllite #	Locality re: Steenbok Shear	Grid ref. Annexure 1
PB592*	x	x	x	x	x	x									west	C 9
PB317	x		x	x	x	x	x	x							east	Q 4
PB510	x		x		x			x			x	x	x	x	east	Q 4
PB511	x	x	x	x	x	x	x	x							east	Q 4
PB578	x	x				x	x		x	x					east	P 4

* near Penssleep beacon (No 70)

\# replaced by sericite/chlorite

The quartzite-schist horizon can be traced south-eastwards across the area, below narrow outlier prongs of Nama Group cover, west of T'Gabies. Here its continuity is terminated, as no quartzites are shown on Theart's (1980) map at this stratigraphic position farther eastwards. The quartzites below Theart's (1980) muscovite gneiss unit (map code mg) appear tectonically discordant to the underlying rocks on their southern side, as is the case for the Ratelfontein - Chabiesies zone west of 17°30'. Theart (1980), however, groups all quartzitic rocks (map code mq) with his "Rietkloof Formation", most occurring stratigraphically just below his "Transitional Zone" (cf. Fig. 3.6).

Some of the lithologies of the Ratelfontein suite correspond with those of the Groothoek Formation (Blignault et al., 1983). However, there are two reasons for giving this formation a new name: firstly a regional correlation with the Groothoek Formation has not yet been proved, and secondly, the quartzite/schist association of the presently defined Ratelfontein suite is the only reliable marker unit for field mapping.

In the study area the Ratelfontein is separated from the Chabies-ies suite by the Kouefontein granite gneiss and Sabieboomrante adamellite gneiss, both of which here have a very broad outcrop.

In the area east of 17°30′ the Kouefontein granite gneiss appears greatly reduced in thickness. Here Theart (1980) includes it ("muscovite gneiss"; map symbol mg) in the Rietkloof Formation. Blignault et al. (1983) refer to this rock as a leucocratic granite and allocate it to the Vioolsdrif Suite (Fig. 3.6).

The Sabieboomrante and Kouefontein granitoids separate formations of divergent orientation on their northern and southern sides. This implies that the base of the Ratelfontein suite is an important zone of lithological discontinuity and structural discordance, either a major unconformity or fault, or both.

3.1.3 Groothoek suite (Richtersveld Subprovince)

Following conformably/concordantly on the Ratelfontein suite with a predominant north-westerly strike and moderate dip to the north-east, this suite is composed of semi-pelitic rocks and a few biotite gneiss layers. Lithologically it is not as diverse as the underlying Ratelfontein suite, but is characterized by abundant conformable pegmatites and granitic gneiss, more so than any other "formation" in the area. It has a broader outcrop between the Tierkloof and Steenbok Shears, where it is about 1700 m thick. East of Steenbok Shear its outcrop is not as wide, while west of Tierkloof Shear it pinches out along strike (cf. Fig. 3.1 and Annexure 1). Its three-dimensional geometrical configuration is wedge-shaped, apparently tapering to the south and west (see Fig. 3.11).

3.1.3.1 Lithologies and Field Description

Muscovite-quartz schist

These rocks are ochre to brown weathering, well foliated and locally crenulated. A quartz rodding averaging 1 cm diameter is present and imparts a characteristic fabric to these semi-pelites. The prominent rodding lineation plunges down dip.

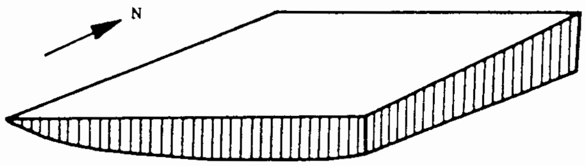


Fig. 3.11 Schematic diagram showing proposed three-dimensional shape of the Groothoek suite. Note that strata taper to the south and west. **This configuration drawn in the initial pre-strained state.**

These schists are most thickly developed between the Steenbok and Tierkloof Shears. West of Tierkloof Shear they occur as lenticular units, conformable within the leucocratic gneisses of the Groenrivier suite, but do not extend as far as the Kromnek Shear.

A petrographic description is given in Table 3.10.

TABLE 3.10 Petrographic Description of mu-qtz schist of the Groothoek suite
Texture: Variable inequigranular to granular texture.
Quartz: Inequant oval-shaped grains, sometimes show marked undulose extinction. Sutured boundaries between grains a characteristic feature.
Muscovite: Forms about one third of the rock; imparts a characteristic fabric expressed either as a sigmoidal cleavage, or a crenulation cleavage, especially near the top of the sequence. Thin muscovite laths aligned along the cleavage form folia spaced some 2 mm apart.
Iron ore: Makes up approximately 5% of the rock in the form of anhedral to euhedral small grains.
Accessories: Plagioclase and epidote - the latter associated with phyllosilicate layers.

Muscovite-quartz-chlorite-sericite schist

This occurs immediately above, and conformable with the Ratelfontein suite east of Kouefontein Shear. Chlorite and sericite give the rock a greenish outcrop appearance, and define a zone of significant retrograde metamorphism.

A petrographic description is given in Table 3.11.

TABLE 3.11 Petrographic Description of mu-qtz-chl-ser schist, Groothoek suite
Texture: Inequigranular rock; extreme range in grain size.
Quartz: Inequant shape, aligned parallel to the foliation. Some larger grains have undulose extinction and grain boundary sutures. Often displays mortar texture, with small quartz subgrains surrounding large grains. In some thin sections quartz is disrupted by sericite along fractures, giving rock a characteristic "web texture".
Muscovite: Has two habits: (i) short broad laths which form decussate clusters, or sometimes fairly large anhedral porphyroblasts; (ii) long slender laths forming folia, 0,5 mm thick, and cutting across the modified granoblastic-polygonal texture of the rock.
Chlorite: Forms fairly broad subhedral blades, completely replacing pre-existing biotite.
Sericite: Occurs along the foliation, locally in broad bands up to 10 mm wide where it has replaced all pre-existing minerals. It can make up to 30% of the rock.
Magnetite: Euhedral to anhedral crystals up to 2 mm in size, generally associated with muscovite and sericite, and represent a late stage phase. Sometimes appear to be drawn out along the foliation.

Biotite gneiss

These fine-grained rocks occur as thin metavolcanic horizons, at most some 2 m thick, within this predominantly metasedimentary sequence. They were located only east of Kouefontein Shear where they appear at the base, and slightly higher up in this formation.

In outcrop they have a greenish colour, with a very strong down-dip lineation. Composed essentially of quartz and biotite with some plagioclase, accessories include epidote, iron ore, muscovite, chlorite and sericite.

A petrographic description is given in Table 3.12.

TABLE 3.12 Petrographic Description of bt gneiss, Groothoek suite

Texture: Equigranular; granoblastic quartz and plagioclase groundmass is transected by folia of olive-green to brown biotite which imparts a sigmoidal cleavage to the fabric.

Quartz: Present as inequant polygonal grains (average size 0,5 mm), but also as larger grain clusters where individually can be up to 2 mm size. The latter generally show marked undulose extinction, with suturing between the grains, and may be flanked by recrystallized polygonal grains (average size 0,1 mm).

Plagioclase: Andesine (optically determined), occurs in subhedral to euhedral laths (up to 1,5 mm long); makes up between 20-40% of the rock. Most show some degree of saussuritization.

Biotite: Olive-green to brown colour, in distinctive parallel clusters of long slender and broad laths <1 mm long, grows across the quartz plagioclase granoblastic-polygonal texture. Biotite clearly defines a distinctive sigmoidal-shaped foliation, in some areas leaving small augen of quartz and plagioclase devoid of phyllosilicates. Individual biotites show undulose extinction due to kinking.

Epidote: Anhedral grains (size < 0,7 mm) associated mainly with biotite, but locally incorporated in the granoblastic-polygonal quartz and plagioclase minerals associated with sericite.

Muscovite: A minute amount present as irregularly shaped laths, mostly < 0,1 mm long. Euhedral to subhedral magnetite <1 mm in size are a late mineral phase.

Chlorite: Replaces biotite and sericite, but is present only in minor proportions.

Mineral assemblages of schistose rocks of the Groothoek suite are summarized in Table 3.13.

TABLE 3.13 Mineral assemblages of schistose rocks, Groothoek suite.

Sample No.	quartz	plagioclase	muscovite	biotite	sericite	epidote	iron ore	chlorite	Locality re: Steenbok Shear	Grid ref. Annexure 1
PB 17	x		x			x			west	K 5
PB 22	x		x			x			west	J 6
PB 42	x	x	x			x	x		west	L 7
PB294	x		x		x		x	x	east	Q 4
PB298	x	x	x	x	x	x	x	x	east	P 4

3.1.3.2 Correlation

These predominantly semi-pelitic rocks do not outcrop west of Kromnek Shear but extend south-eastwards of 17°30′ and appear to correspond to the lower portion of the "Transitional Zone" (Theart, 1980; map codes sg(1), mbg; see Fig. 3.6). The presence of sericite, particularly characteristic of this zone in the study area, is also noted by Theart (1980, p.11). The mainly schistose lithologies of this formation in the present area are similar to those occurring in the Groothoek Formation (Blignault et al., 1983). They are, however, dextrally offset along a fault on T′Gabies just east of 17°30′, but are stratigraphically in the correct position to justify correlation with similar strata in the present area.

3.1.4 Groenrivier suite (Richtersveld Subprovince)

This, the upper part of the transition zone of Ritter (1980, p. 70), is mostly conformable with the Groothoek suite, but differs significantly from it in the following characteristics:

- extremely diverse supracrustals, including leucocratic and mesocratic gneisses, semipelitic and pelitic schists;
- lateral lithological variability;
- intercalation of felsic to ultramafic plutonic rocks, especially at the top of the sequence;
- widely varying north-easterly dip-angles to the foliation.

Its top is defined as the northernmost contact with continuously outcropping Vioolsdrif granodiorite, which in this area coincides with the Groothoek Thrust zone. It has a maximum thickness of about 1200 m but tectonic thickening probably accounts for part of this. Plutonic rocks constitute at least one third of the suite.

The average strike direction is north-westerly, with a variable north-easterly dip, slightly different from the Ratelfontein suite to the south where the strike is more west-northwesterly. East of Kouefontein Shear this angular discordance is expressed as a distinct triangular shape of the intervening Groothoek suite (see Annexure 1, grid reference 03).

Semipelitic muscovite and muscovite biotite schists generally constitute the major rock-types occurring both as intercalations within the gneisses and as discrete large lenticular units, especially in the north-eastern part of the area. Quartzo-feldspathic gneisses form minor thin bands, as do the pelitic units within the sequence.

3.1.4.1 Lithologies and field description

3.1.4.1.1 Semi-pelitic rock-types

Muscovite schists

Occurring mainly in the eastern part, these rocks display a characteristically well-developed platy foliation and down-dip stretching lineation.

A petrographic description is given in Table 3.14.

Biotite-quartz schist

Biotite-quartz schists are commonly interbedded with most lithologies of this suite. A petrographic description is given in Table 3.15.

TABLE 3.14 Petrographic Description of mu-schist, Groenrivier suite
Texture: Granular, Granoblastic-polygonal texture, muscovite defines the foliation.
Quartz: Shows extremely variable grain-size, small polygonal grains either form clusters or surround large grains in an incipient mortar texture. Sutured boundaries present.
Plagioclase: (oligoclase) forms mostly anhedral grains (<1,5 mm),noticeably saussuritized.
Muscovite: Occurs as slender euhedral laths, <1 mm long, interleaved and clustered together in 0,5 mm-thick folia. Transects rock in a sigmoidal cleavage (Plate 3). Minor proportion of muscovite present as broad, randomly oriented euhedral laths.
Accessories: Minor anhedral epidote and olive-green subhedral biotite. The latter occasionally interleaved with muscovite. Magnetite makes up less than 3% of the rock.

TABLE 3.15 Petrographic Description of bt-qtz schist
Texture: Granoblastic-polygonal, characterized by a very pronounced sigmoidal cleavage, defined by biotite.
Quartz: Generally varies from 0,5 to 1 mm in size, occasional clusters or individual grains may reach about 3 mm.Many grains inequant, aligned parallel to the foliation. Large ones commonly display pronounced undulose extinction and sutured grain boundaries. Small polygonal quartz grains 0,1 mm in size occur in narrow zones < 0,2 mm thick, parallel to the foliation.
Plagioclase: (oligoclase) forms small subhedral laths averaging 0,3 mm in size, generally showing advanced stages of saussuritization.
Biotite: Olive-green laths very pronounced; account for 25% of this rock. Parallel clustering of laths develops folia, averaging 0,5 mm thick and 1 to 2 mm apart, continuous and conspicuous across entire thin section. Incipient replacement of biotite by chlorite evident.
Accessories: Apatite and epidote; mostly as anhedral grains of variable size, notably associated with biotite. Iron ore as anhedral grains, largely replaced by chlorite. Stringers of sericite between polygonal grains of quartz and plagioclase.

3.1.4.1.2 Pelitic rock-types

Phlogopite-corundum schist

Near the base of this suite a greenish, very coarse-grained, lenticular phl-cd schist unit up to 7 m thick crops out discontinuously along strike for at least 4 km (see Annexure 1, grid reference O3 and P3, and Annexure 2). Its orientation is approximately the same as the enclosing beds.

This rock is composed predominantly of phlogopite with lesser proportions of corundum, sericite and accessory sphene and muscovite (Fig. 3.12). A petrographic description is given in Table 3.16.

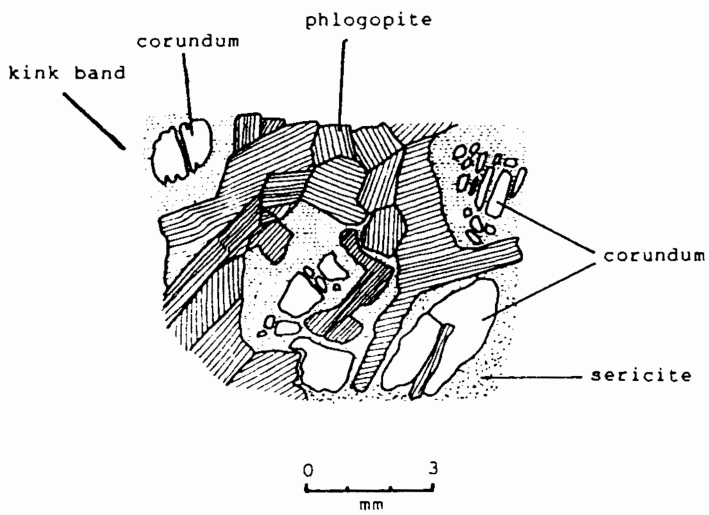


Fig. 3.12 Textures of a phlogopite-corundum schist, Groenrivier suite. Note advanced replacement of corundum porphyroblasts by sericite. Drawn from a photomicrograph.

This unit extends westward only as far as the Tierkloof Shear (grid reference K5, Annexure 1), where it is cut off by the Groothoek Thrust zone. Here large tourmaline crystals are associated with this rock-type. X-ray determination revealed them to be dravite (composition: $Na\ Mg_3\ Al_6\ B_3\ Si_6\ O_{27}\ (OH)_4$; see Appendix A-1.4).

TABLE 3.16 Petrographic Description of phlogopite-corundum schist of the Groenrivier suite
Phlogopite: Generally shows a preferred orientation. Occurs as slightly pleochroic broad blades (average dimensions 5 x 1 mm). Some arranged in a decussate texture, disrupting the parallel fabric in the rock. Most are bent to a slight degree, exhibiting undulose extinction. Occasional very narrow kink bands traverse the rock at a steep angle to the predominant schistosity (Fig 3.12).   **Corundum:** Occurs mostly as large sporadic porphyroblasts, surrounded by a replacement rim of sericite. Size of remnant crystals is extremely variable, depending on the degree of replacement. Sometimes entire grains have been replaced, but not unusual to find remnant grains displaying their characteristic parting.   **Sericite:** Can make up about 15% of this rock.   **Accessories:** Sphene as smallish (< 0,3 mm) brown grains; and muscovite occurring as small (average 0,1 mm) stubby laths, mostly along the edge of the sericite replacement rim.

A geochemical analysis showed high abundances of MgO and Al_2O_3 (cf. Table 3.6, sample PB 499), consistent with the mineralogical preponderance of phlogopite and corundum. There are some analogies between the geochemical composition of this unusual rock and that of altered tuffaceous rocks in the Green Tuff region of Japan (see sample #2, Table 3.6). The Groenrivier phl-cd metapelite is, however, enriched with respect to alumina and potash, and depleted in FeO, Na_2O and CaO, compared to the Japanese example. If these rock-types were originally comparable, then the elements Fe, Na and Ca could have been removed during weathering of the Groenrivier rock. Alternatively, characteristically high K, Mg and the presence of dravite strongly suggests that its precursor may have been an evaporite (Moine et al., 1981).

Cordierite schists

Generally less than 1 m thick, these occur together with the phl-cd schist. They are found as discontinuous lenses at its base, also as thin lenticles throughout the remaining sequence, except in the eastern part of the area adjacent to the Riethoek Shear (grid reference R4, Annexure 1). Here the crd-schist occurs stratigraphically below a prominent amphibolite unit, is up to 20 m thick, but wedges out rapidly westwards.

These schists crop out as unmistakable units in which prominent cordierite porphyroblasts protrude as cigar-shaped grains, reaching a maximum size of 15 x 5 cm (Plate 4). These linear structures plunge predominantly towards the north-east at a consistent angle.

Cordierite porphyroblasts are completely replaced by sericite, whereas the groundmass of quartz, chlorite and sericite shows a grossly inequigranular texture. Quartz shows an extreme size range, the larger grains occurring usually in clusters, exhibiting undulose extinction. Olive-green biotite occurs as small laths within broader bands 7 to 8 mm apart, imparting a distinct foliation. Chlorite is present largely as a replacement of biotite and iron ore, only portions of the latter mineral still remaining.

A geochemical analysis revealed a high SiO_2 content, and moderate abundances of Al_2O_3 and K_2O (cf. Table 3.6, sample PB 500). Compared with other metapelites from the Bushmanland Subprovince (e.g. sample #1, Table 3.6) this analysis reflects average geochemical values (Moore, 1986). Similar pelitic rock types are described by Ward (1977) in an area to the east of 17°30' where their mineral paragenetic diversity increases towards the east. Cordierite schist horizons, however, thin out rapidly westwards, and within the study area the size of porphyroblasts is not as great as the 25cm length described by Ward (1977).

3.1.4.1.3 Quartzo-feldspathic and hornblende gneisses

Occurring throughout this sequence as thin discontinuous layers, especially in the area east of Kouefontein Shear, these characterize the Groenrivier suite, albeit only in minor proportions. Between the Steenbok and Tierkloof Shears they are entirely missing, but reappear west of the latter shear.

Geochemical analyses reflect typically high SiO_2, Al_2O_3 and CaO associated with silicic volcanic rocks (cf. Ritter, 1980).

Quartzo-feldspathic gneiss

Thin (usually < 1 m thick), discontinuous layers of cream to pinkish, massive quartzo-feldspathic gneisses occur especially near the base of the Groenrivier suite, intercalated with the schistose rocks. They are composed of quartz, plagioclase and microcline, with minor microcline-microperthite, muscovite, biotite, epidote, iron ore, calcite and chlorite.

A petrographic description is given in Table 3.17.

Table 3.17 Petrographic Description of quartzo-feldspathic gneiss of the Groenrivier suite

Texture: Fine-grained (average size 0,5 mm) rock;inequigranular texture. The foliation is defined by elongate thin laths of muscovite and olive-green biotite with a preferred dimensional orientation.

Quartz: Anhedral quartz grains range in size from 0,5 to 1 mm. Larger grains show undulose extinction.

Feldspar: Anhedral to subhedral microcline (up to 2 mm long) and minor microcline-microperthite are also present. Plagioclase shows advanced saussuritization.

Phyllosilicates: Muscovite and biotite associated with bands of smaller polygonal quartz and saussuritized plagioclase grains. Biotite shows replacement by chlorite.

Accessories: Small anhedral epidote grains and anhedral calcite grains. Magnetite forms very small (usually <0,2mm) euhedral grains and is clearly a late stage crystallization phase.

Hornblende gneiss

Thin discontinuous pale-green hornblende gneisses occur sporadically throughout the Groenrivier suite. Their massive character results in positive-weathering rounded-boulder morphology. They are not reliable markers because of their impersistence along strike, and the fact that they are disrupted by the north- northeast trending Pan-African shear zones in this area.

Fine-grained and equigranular, the rock consists mainly of plagioclase and quartz in a granoblastic-polygonal texture, with sporadic blue-green hornblende. The granoblastic-polygonal texture is outlined by an interleaving network of very small sericite grains. Accessory minerals include brown biotite, epidote and iron ore. Chlorite has replaced some biotite and amphibole only marginally.

Where hornblende is more abundant it occasionally develops a garbenschiefer texture showing elongate prismatic-shaped grains in a random orientation throughout the rock (Plate 5). The rock can be derived either from impure calcareous sediments, or igneous rocks such as tuffs, intermediate to basic varieties (Spry 1969, p. 269). The garbenschiefer texture is in this case probably due to growth of hornblende along the foliation during the post-tectonic phase of Namaqua metamorphism (Chapter 5).

Hornblende-gneiss which differs considerably from other occurrences is found at the top of the suite, about 1,5 km south-west of the Bluff beacon (number 44), in the north-east of the area. This unit, which outcrops sporadically along strike for some 4 km, has prominent porphyroblasts of randomly oriented hornblende in a strongly foliated quartzo-feldspathic groundmass (Plate 6).

Mineral assemblages of schistose and gneissic rocks of the Groenrivier suite are summarized in Table 3.18.

TABLE 3.18 Mineral assemblages of schistose and gneissic rocks, Groenrivier suite.

Sample No.	quartz	plagioclase	muscovite	biotite	hornblende	epidote	sericite	chlorite	apatite	iron ore	phlogopite	corundum	sphene	tourmaline	Locality re: Steenbok Shear	Grid ref. Annexure 1
PB605			x				x				x	x	x	x	west	K 5
PB218			x				x				x	x	x		east	P 3
PB266	x	x		x			x	x		x					east	R 4
PB267	x	x		x		x	x	x	x	x					east	R 4
PB292	x	x		x	x	x		x		x					east	P 3

3.1.4.2 Correlation

The hornblende-gneisses of the Groenrivier suite probably represent silicic to intermediate metavolcanic horizons. They can be correlated with hornblende gneisses from east of the study area, where they are spatially confined to the southern margin of the Vioolsdrif batholith (Ward, 1977).

Nodular cordierite schists outcrop more extensively east and south-east of the present area (Ward, 1977; Theart, 1980; Blignault et al., 1983), and are there associated with a greater variety of pelitic and volcanic rock-types. The Groenrivier suite correlates with the upper portion of the "Transitional Zone" of Theart (1980, p. 10; and cf. Fig. 3.6), based on the presence of cordierite-bearing rocks characteristic of the "Transitional Zone". Blignault et al. (1983), however, include nodular cordierite schists occurring in the Namaqua geotraverse, in the Tsams/Hom Formation of the Haib Subgroup. In the study area the nodular cordierite schists of the Groenrivier suite are of uncertain stratigraphic position, but can probably be correlated with the Groothoek Formation (Blignault et al., 1983).

3.1.5 Windvlakte suite (Richtersveld Subprovince)

Ritter (1980, p. 70) classifies all metavolcanics occurring south of the Rosyntjieberge as belonging to the Windvlakte Formation, and noted that they are more recrystallized in the southern outcrop area than metavolcanics north of the Rosyntjieberge. The meta-volcanics have been intruded and disrupted by Vioolsdrif granodiorite, making it likely that different stratigraphic levels in the sequence are exposed at the present erosion surface. The Windvlakte metavolcanics are geochemically similar to the granodiorite plutonic intrusives (Ritter, 1980, p. 71). This geochemical kinship between extrusive and plutonic rocks has been corroborated by Reid (1979a) for the area straddling the lower Orange River.

In the latter area, Reid (1977) shows that a differentiated calc-alkaline suite of metavolcanic rocks is present. These lavas, of which meta-andesite is the most abundant type, were erupted 2000 Ma ago. Reid (1977) proposes that the younger lavas were derived by fractional crystallization from a basaltic andesite parent, whereas the earlier rhyolitic lavas probably originate from a separate crustally-derived magma.

In the present area Windvlakte rocks occur as xenolithic fragments of variable size within the Vioolsdrif granodiorite. Their original geographic distribution has been modified through Late Proterozoic (Pan-African) tectonism. At present they are exposed in three distinct areas (see Fig. 3.1, and Annexure 1):
- the north-west, just west of Kromnek Shear;
- the centre-north area, between Tierkloof and Steenbok Shears;
- the extreme north-east, to the west and south of Bluff beacon.

For descriptive purposes rock-types of this suite are divided into upper and lower units. The relationship between them is not clear, as the lower unit occurs only east of Steenbok Shear, and the upper unit only west thereof.

3.1.5.1 Lithologies and field description

3.1.5.1.1 Upper Unit

Four rock-types occur in the area between Steenbok and Tierkloof Shears.

Quartzo-feldspathic gneiss (mesocratic)

A mesocratic, medium to dark grey, fine-grained, recrystallized quartzo-feldspathic gneiss is the most abundant rock-type in the western and eastern parts of this central area. It has a massive unfoliated field appearance. In thin section recrystallized phenocrysts of highly saussuritized plagioclase sporadically occur within a recrystallized ground-mass of quartz, plagioclase, microcline, epidote and calcite. Small biotite grains are almost completely replaced by chlorite. Muscovite forms anhedral grains or small euhedral laths. Anhedral epidote is associated with both the phyllosilicates and plagioclase.

Two deviations from the predominantly massive texture in these rocks are noted:
- an outcrop at Kliphoogte where an original bedding (S_o) can be identified in small metavolcanic xenolithic blocks in the Vioolsdrif granodiorite (grid reference L4). Defined by a compositional banding, it probably represents a fine-grained original layering in bedded tuff;
- west of the road to Eksteenfontein (Grid Reference K2; Annexure 1) an outcrop shows a transposed foliation in the meta-volcanics. Here leucocratic lenses 0,5 cm thick and some 2 cm long have been disrupted by a transgressive biotite foliation in an otherwise dark grey massive siliceous rock (Plate 7).

This fine-grained rock with a modified granoblastic polygonal texture is composed of quartz, plagioclase, biotite, muscovite, epidote, calcite and chlorite. Quartz, plagioclase and muscovite form thin bands which in places are transected by small olive-green biotite laths outlining ovoid-shaped leucocratic units up to 5 mm in length. Occasional clusters of quartz grains are present. Both biotite and muscovite occur as short subhedral to euhedral laths showing undulose extinction. Anhedral epidote and calcite are fairly abundant.

Quartzo-feldspathic gneiss (leucocratic)

Towards the central area a pale pink to greenish, fine-grained quartzo-feldspathic gneiss appears in a zone trending approximately north-south. It is composed predominantly of quartz, plagioclase and muscovite, with epidote in thin transgressive veins. A similar rock-type occurs towards the southern fringes of their area of outcrop where it often has a paler colour in the field. These rocks form xenolithic units within the Vioolsdrif Granodiorite and are therefore not very extensive.

Porphyritic quartz-feldspar gneiss

Immediately flanking the eastern contact of the previous rock-type is a zone a few tens of metres wide containing a whitish, coarse-grained rock, with large quartz phenocrysts. Where the Pan-African foliation ($S(P)_2$) transects this rock, it has a matrix of quartz, feldspar and chlorite surrounding the phenocrysts.

Banded Iron Formation

This rock-type was located approximately 2 km east of Kliphoogte. Blackish and composed predominantly of quartz and magnetite, it occurs as pods or lenses 0,5 to 1 m thick and only a few metres long, in the grey recrystallized quartzo-feldspathic gneiss.

Just above the Windvlakte Thrust (grid reference L2, Annexure 1), the metavolcanics include dark-green porphyritic varieties slightly coarser-grained and less equigranular than those to the south. Phenocrysts are made up of augen-shaped clusters of larger quartz, and/or plagioclase and microcline grains, separated by a matrix of plagioclase, quartz, microcline, actinolite, epidote with lesser zircon and apatite grains. Heavily saussuritized plagioclase forms anhedral porphyroblasts up to 2 mm in size. Quartz grains have undulose extinction, with suturing between the grains.

Short subhedral olive-green biotite laths outline the foliation in this rock. Randomly oriented, pale-green actinolite generally forms prismatic grains, usually no longer than 2 mm. It sometimes has an elongate radiating habit where it forms subhedral grains. Anhedral grains of epidote of variable size occur spread throughout the thin section.

West of Kromnek Shear gneisses of the Windvlakte suite contain a tectonised fabric. They are dominantly a mesocratic variety, similar to those north of Kliphoogte. The metavolcanic rocks in this western sector are essentially inequigranular. Large inequant quartz and plagioclase grains occur in clusters, the former showing undulose extinction. Plagioclase is saussuritized. A modified granoblastic-polygonal texture is common, and deformational fabrics are pronounced. Muscovite laths, for instance, strongly outline the foliation in the vicinity of a broad thrust near the southern contact with the Tweeriviere granite. Within the thrust zone, they form distinct "fish" shapes in a mylonitic rock. A few thin sections show green biotite associated with anhedral epidote grains and euhedral magnetite. Chlorite is present as a replacement of biotite, and as discrete anhedral grains.

3.1.5.1.2 Lower Unit

In the eastern sector just west of the Bluff beacon xenolithic rafts of quartzo-feldspathic gneisses crop out within the Vioolsdrif granodiorite, just north of the Groothoek Thrust. They differ from those south of the Groothoek Thrust in being more massive and more abundant, with the pelitic component being virtually absent.

Quartzo-feldspathic gneiss

Fine- to medium-grained, beige coloured, massive rocks, composed essentially of quartz, plagioclase, muscovite and epidote, they develop a characteristic hollowed-out honeycomb weathering in places on outcrop. Various forms of these siliceous rocks, including a more porphyritic variety, are the predominant rock-type here.

Hornblende gneisses

Minor components of these fine-grained rocks resemble some of the thin bands occurring in the Groenrivier suite. Quartz and plagioclase form a granoblastic-polygonal texture, the plagioclase in places displaying advanced saussuritization. Small muscovites and epidotes sometimes decorate the boundaries between other polygonal grains.

Blue-green hornblende forms typical long subhedral prismatic crystals up to 6 mm long in random orientation, displaying the typical "bow-tie" texture in places. This garbenschiefer texture is a post-tectonic crystallization feature associated with the Namaqua metamorphic event (see Chapter 5).

Epidote is fairly abundant, sometimes along a foliation if this is developed, and at other times filling veins. It completely replaces hornblende in places. A minor amount of apatite is also present. Anhedral grains of iron ore form a small proportion of the rock by volume.

Mineral assemblages of metavolcanic rocks of the Windvlakte suite are summarized in Table 3.19.

3.1.5.2 Correlation

The upper unit of the Windvlakte suite consists mainly of metavolcanic rocks, with minor bedded tuff horizons and associated chemical sediments. Although Ritter (1980) originally classified the Windvlakte Formation on a geographic rather than a petrographic basis, its correlation with other supracrustal units of the Richtersveld Subprovince is a topic beyond the scope of the present investigation. However, in view of their geographic posi-

TABLE 3.19 Mineral assemblages of metavolcanic rocks, Windvlakte suite.

Sample No.	quartz	plagioclase	microcline	hornblende	muscovite	biotite	epidote	actinolite	chlorite	calcite	apatite	iron ore	Locality re: Steenbok Shear	Grid ref. Annexure 1
PB352	x	x	x		x	x	x		x			x	west	G 2
PB354	x	x			x	x	x		x	x			west	K 4
PB482	x	x	x			x	x	x			x		west	L 1
PB230*	x	x		x	x		x					x	east	Q 2

* North of Groothoek Thrust

tion, the possibility exists that these metavolcanics represent a localized manifestation of igneous activity, situated well to the south of the main lava outpourings that comprise the Haib Subgroup. In this study they are correlated with the Haib Subgroup.

Some members of the lower Windvlakte unit, e.g., the leucocratic quartzo-feldspathic gneisses, can probably be correlated with those described by Ward (1977) as "leptite" and by De Villiers and Söhnge (1959) as the "grey granulite". Ward (1977) regards the leptite as a marginal correlative of the Haib Subgroup.

3.1.6 Correlation of supracrustal rocks, and their original geologic setting

Supracrustal rocks in the area can be correlated with those to the east of 17°30′ where similar rocks are described by Theart (1980), Ward (1977) and Blignault et al. (1983). The Chabiesies suite appears to correlate with the major part of the Rietkloof Formation (Theart, 1980), and corresponds to the Kabına Formation (Blignault et al., 1983), whereas the Ratelfontein suite probably correlates with the upper metaquartzite of the Rietkloof Formation (Fig. 3.6).

Supracrustal units form laterally inextensive and discontinuous units. Underlying a large part of the southern area, the Chabiesies suite in the west strikes almost orthogonally to the formations immediately north of it. In the south-east its lithological and structural trends diverge obliquely eastward from those in the overlying formations. Eastwards for some 25 km beyond 17°30′ there is a consistent similar pattern, as highlighted by Theart's (1980, p. 52) structural study, suggesting that a major unconformity and/ or tectonic discordance exists at or near the top of the Chabiesies suite. This zone of discordance coincides with the position of the leucocratic granite (Fig. 3.6, Mvl map symbol of Blignault et al., 1983), although the discordance is more noticeable farther towards the south-east.

The Ratelfontein suite has a north-westerly trend, outcropping across the entire area. Although lithologies are offset along Pan-African shear zones this suite is a valuable marker unit. Its consistent trend over most of the area contrasts markedly with the variable lithological trends in the Chabiesies suite to the south.

In the study area the Chabiesies and Ratelfontein suites are separated by the very thick Kouefontein granite gneiss sheet (Annexure 1). Eastwards of the 17°30′ boundary these granitic rocks (map code mg: Theart, 1980; and Mvl: Blignault et al., 1983) occupy a much reduced outcrop area eventually tapering out about 25 km farther on. Theart (1980) includes these as paragneisses in his "Rietkloof Formation", but in the present area the Kouefontein body shows obvious orthogneissic features which support an originally intrusive plutonic origin (see section 3.2.10).

The Groothoek suite has a westward tapering wedge configuration, extending only a short distance west of Tierkloof Shear. It can be traced eastwards as the lower portion of the "Transitional Zone" where metapelites containing cordierite-anthophyllite assemblages become far more abundant. A lenticular lithological pattern within the suite is strongly suggestive of considerable internal imbrication and a tectonic origin for this unit.

The Groenrivier suite is characterized by facies changes in semipelitic rocks, and discontinuous metavolcanic beds. This probably reflects deposition accompanied by intermittent volcanic extrusion.

Distribution of supracrustal units in the southern Richtersveld shows that there was widespread volcanism during the Early Proterozoic. Metavolcanics of the Windvlakte suite are extensively developed across the entire southern area of the Richtersveld. The lower unit consists of leucocratic and mesocratic fine-grained quartz feldspar gneisses, but there is a greater diversity of volcanic rock-types in the upper unit. Furthermore the grade of metamorphism is slightly lower in the upper unit.

There is still a problem about the substratum or basement rocks to the upper unit volcanics, for none have yet been found. Thin metavolcanic beds in the lower unit of the Windvlakte suite, occurring as rafts in Vioolsdrif granodiorite, are similar in field appearance to those in the Groenrivier suite ("Transitional zone ").

From field observations I propose that the lower unit of the Windvlakte suite originally formed a continuous succession with the Groenrivier, Groothoek and Ratelfontein suites, if granitic sheets which now separate them are ignored. This essentially suggests that volcanic rocks of the Richtersveld Subprovince form part of the same package as gneisses and schists immediately to their south. Ritter (1980) interpreted the latter as part of the Bushmanland Subprovince, lithologies of both subprovinces forming one conformable sequence. However, my interpretation of the spatial relationship between lithologies of the Richtersveld Subprovince and those of the Bushmanland Subprovince differs from that of Ritter (1980) by the following reasoning:

- rocks of the Groenrivier, Groothoek and Ratelfontein suites do not constitute part of the Bushmanland Subprovince because they have a close association with volcano/ plutonic lithologies of the Richtersveld Subprovince. I speculate that this spatial relationship points to an island arc origin for these rocks (either fore-arc or back-arc basin deposits, Chapter 6);
- the north-westerly strike orientation of the Groenrivier, Groothoek and Ratelfontein suites in the study area and east of 17°30′, contrasts strongly with the divergent lithological orientation of the Chabiesies Suite in the south. It may represent an original regional unconformity or a zone of tectonic dislocation. Similarly, an unconformable relationship is suggested from the orientation of strata along this zone in the area farther east (Theart, 1980). In the study area, however, intensely tectonised strata of the Ratelfontein suite show that dislocation occurred at the base of this suite, prior to intrusion of the Sabieboomrante and Kouefontein granitic sheets (section 3.5.3; and Chapter 4). I speculate that lithologies of the Bushmanland Subprovince (Chabiesies suite) formed as continental shelf deposits because of their widespread distribution in the study area and farther eastwards.

I suggest therefore that the zone separating lithologies of the Richtersveld Subprovince from those of the Bushmanland Subprovince lies at the base of the Ratelfontein suite. Field evidence strongly suggests that it can be interpreted as a significant zone of dislocation. There is therefore no doubt that the boundary between the Richtersveld and Bushmanland Subprovinces is a tectonic one. This interpretation of the relationship between the two subprovinces differs markedly from that of Ritter (1980) who regarded this boundary as a conformable one.

3.2 EARLY TO MID-PROTEROZOIC INTRUSIVE ROCKS

3.2.1 Vioolsdrif Granodiorite (V.G.)

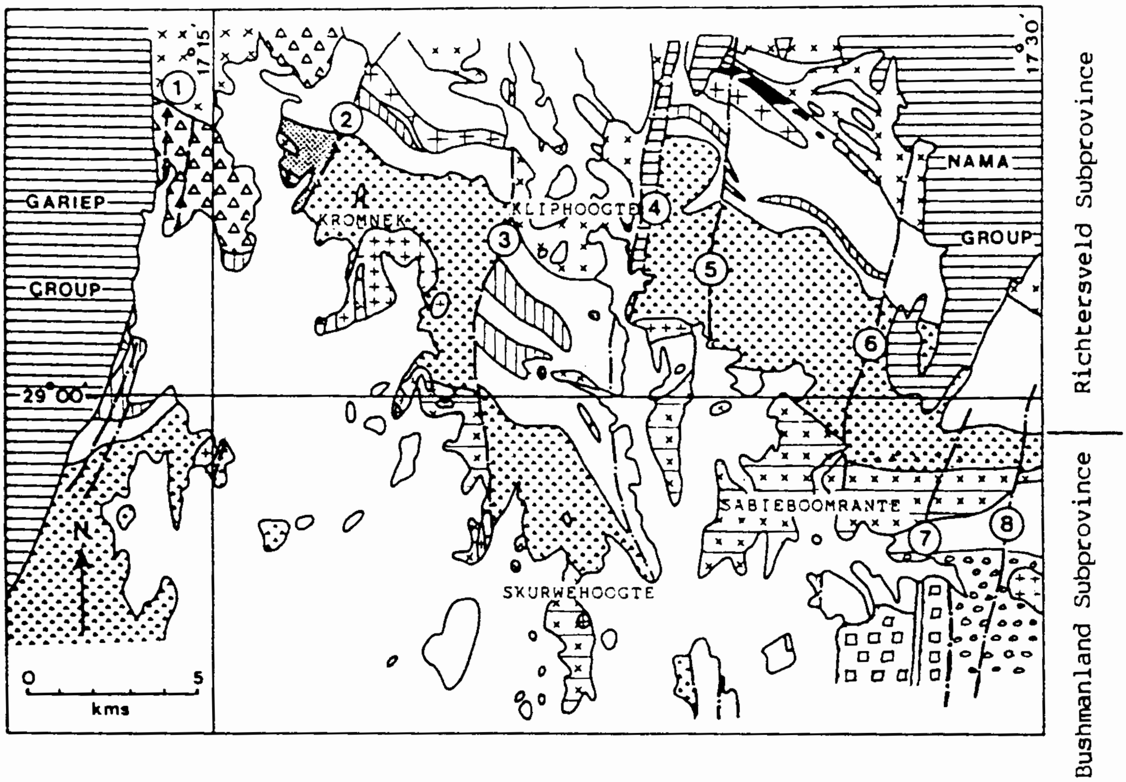
The Vioolsdrif Suite comprises ultramafic/mafic, diorite, tonalite granodiorite, adamellite and leucogranite rocks taking the form of a composite batholith and occupying an area of at least 30,000^2 km (Rogers, 1915; De Villiers and Söhnge, 1959; Middlemost, 1963; De Villiers and Burger, 1967; Ward, 1977: Blignault, 1977; Ritter, 1980; and Reid, 1974; 1977; 1979a, b; 1981; 1984). The bulk of intrusive rocks were emplaced soon after eruption of the 2000 Ma Orange River Group lavas, and have been dated at 1900 Ma (Reid 1979b). Reid considers the entire batholith to be a juvenile addition from the upper mantle.

The granodiorite and tonalite make up at least half the batholithic area of exposure (Reid, 1977) and abut the "Transitional Zone" along the entire length of its southern outcrop (see Fig. 3.13 and Annexure 1).

3.2.1.1 Field relationships

The granodiorite ranges from melanocratic to slightly more leucocratic varieties, is easily recognised by its very coarse grain-size, rather spotty nature, and generally greenish-grey colour. Its appearance on weathering varies between a rounded boulder type where the rock is unfoliated, to a slabby, sometimes gneissic nature, especially at its southern foliated contact.

Between the Kouefontein and Riethoek Shears, two characteristics of the Vioolsdrif Granodiorite (V.G.) are prevalent:
- there is a greater abundance of Windvlakte rafts within it nearer the contact zone. This xenolithic characteristic is similar to that of granodiorites in the northern Richtersveld, in contrast to the relatively xenolith-free adamellites and leucogranites (Reid, 1977);
- hornblende occurs at or near its southern tectonized contact, and is confined to east of Steenbok Shear. This characterstic, as well as the platy nature of the granodiorite, is persistent east of 17°30′ (Ward, 1977).



NAMA AND GARIEP GROUPS, Cover rocks
K-GRANITES (SPEKTAKEL SUITE?)
KOUEFONTEIN GRANITE GNEISS
LEUCOGRANITE
QUARTZ FELDSPAR GNEISS - EYAMS GRANITE?
NODULAR GNEISS AND BIOTITE GNEISS - GLADKOP SUITE?
DABBIEPUTS MUSCOVITE BIOTITE GRANITE GNEISS
SABIEBOOMRANTE ADAMELLITE GNEISS
TWEERIVIERE GRANITE GNEISS
MEGACRYSTIC GRANITE
VIOOLSDRIF GRANODIORITE
MAFIC/ULTRAMAFIC ROCKS

VIOOLSDRIF SUITE

1 Stinkfontein Contact Shears
2 Kromnek Shear
3 Tierkloof Shear
4 Steenbok Shear
5 Kouefontein Shear
6 Riethoek Shear
7 Chabiesies Shear
8 Witbakensberg Shear

Fig. 3.13 Intrusive rock-types in the study area. Supracrustals shown unstippled.

Near Bluff beacon in the north-east numerous aplitic veins are present in the granodiorite. These are probably the equivalent of Ward's (1977) "aplogranite" in the area farther to the east.

Biotite-rich selvedges up to several cm long, plunging at moderate to steep angles towards the north, are another characteristic feature of the granodiorites along their southern contact.

3.2.1.2 Petrography

Modal analyses of three samples of this rock-type collected away from the contact zone show that it has a typical granodiorite composition (Table 3.20, and Fig. 3.14). The rock is inequigranular and is composed of quartz, saussuritized plagioclase, K-feldspar, (microcline and microcline-microperthite), with lesser biotite, epidote, and accessory sphene, apatite, iron ore and calcite. Very noticeable is that all the larger quartz grains show undulose extinction, and suturing between them. Green biotite and yellowish epidote usually outline an incipient cleavage in samples north of the contact zone, but define a lepidoblastic texture nearer the contact zone.

3.2.1.3 Geochemistry

Two granodiorite samples in the Kliphoogte area were analysed for comparison with the type area farther north. PB 622 was collected in the Kliphoogte pass and PB 623 some two km to the north-east of the pass. The results are presented in Table 3.21. Major and trace element abundances compare well with other granodiorites from the type area (e.g. sample 2), and a correlation by geochemical inference can be made.

3.2.2 Megacrystic Granite

3.2.2.1 Field relationships

The megacrystic granite crops out in the north western portion of the area, and is overlain in part by the younger Stinkfontein Formation. Described as a porphyroblastic biotite gneiss (De Villiers and Söhnge, 1959, p. 109) and as a megacrystic granite (Ritter, 1980, p. 80), it shows characteristic development of densely spaced subhedral K-feldspar megacrysts up to 5 cm in cross-section. The contact with Vioolsdrif Granodiorite is in places a tectonic one, but elsewhere it is intrusive. During the present survey, one area was located where the contact is distinctly gradational (grid reference D3, Annexure 1). Here feldspar phenocrysts increase in size and frequency over a zone some

TABLE 3.20: Modal composition of Vioolsdrif granodiorite●, Sabieboomrante adamellite gneiss✦ and megacrystic granite◉. Analyses carried out with a Swift Point Counter.

Mineral	PB622●	PB623●	PB606●	PB91✦	PB451✦	PB616✦	PB577◉	PB586◉
Quartz	27	28	28	24	24	32	39	35
Alkali Feldspar	9	20	13	33	29	19	14	9
Plagioclase	38	36	46	31	29	27	29	32
Biotite	21	12	12	10	11	18	12	16
Muscovite	1	2	tr	tr	tr	1	2	4
Chlorite	-	tr	-	tr	1	tr	1	-
Epidote	2	2	1	1	2	2	2	2
Sphene	1	tr	-	tr	1	tr	1	2
Iron ore	-	-	-	tr	1	tr	-	tr
Apatite	tr	tr	tr	-	tr	tr	tr	tr
Calcite	1	-	-	-	tr	tr	tr	tr
Zircon	tr	-	-	tr	tr	tr	tr	tr
Sericite	-	-	-	1	2	tr	tr	tr
TOTAL	100%	100%	100%	100%	100%	99.5%	100%	100%

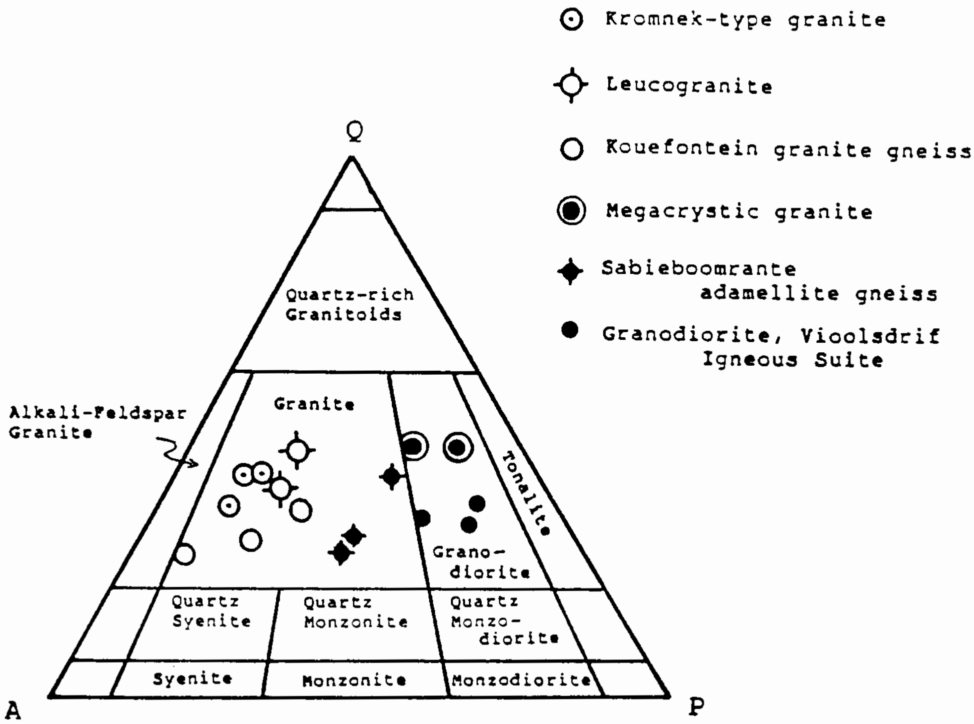


Fig. 3.14 Modal composition of granitic rocks from the southern Richtersveld marginal zone, plotted on a Streckeisen diagram. (Q = quartz, A = alkali feldspar, P = plagioclase)

100 metres wide, when traced from typical Vioolsdrif Granodiorite into the megacrystic granite. The latter therefore appears to be intimately related to the Vioolsdrif Suite.

In the field rare xenoliths of quartzo-feldspathic gneiss and Vioolsdrif Granodiorite-appearing fragments occur in the megacrystic granite, suggesting that it is a slightly younger phase of the Vioolsdrif Suite.

3.2.2.2. Petrography

Modal compositions of this rock show that it falls within the granodiorite field when plotted on a Streckeisen diagram (Table 3.20, and Fig. 3.14).

Tectonized samples of this granite show plastic deformation textures, the megacrysts being deformed into very large augen-shaped glomero-porphyroclasts of quartz, or single microcline grains. These are surrounded by a fine-grained granoblastic-polygonal groundmass of predominantly recrystallized quartz, plagioclase, biotite and epidote. Sometimes quartz forms pronounced ribbons, the grains being mostly recrystallized in contrast to the larger strained grains with marked undulose extinction.

Broad olive-green biotite laths outline the foliation and in places enclose the larger quartz and microcline augen. Abundant anhedral epidote grains are commonly associated with biotite. Sphene makes up 1 to 2% of the minerals present, sometimes forming large rhombic sections. Minor proportions of myrmekite, zircon, apatite, sericite and calcite are also present.

3.2.2.3 Geochemistry

Major element abundances show a SiO_2 content of 61,5%, indicating that the megacrystic granite sample is a more basic rock than typical Vioolsdrif Granodiorite (Table 3.21, sample PB 586). Its composition is closer to that of the tonalites (60% SiO_2). However, texturally and structurally this rock is not typical of the latter. In its trace element geochemistry, Zr, Nb and Y abundances are high compared to typical Vioolsdrif Granodiorite, but Sr in this case is closer to the average values for the latter (see sample 2, Table 3.21). Ba is also slightly anomalous in that, when plotted against SiO_2, it falls outside the field occupied by Vioolsdrif granites (D.L. Reid, personal communication, 1986).

3.2.2.4 Discussion

I propose that the megacrystic granite is intimately related to Vioolsdrif Granodiorite, in view of the field relationships described above. There appear to be some differences in their geochemistry, however, suggesting that the megacrystic granite is more basic in nature than typical Vioolsdrif Granodiorite. The exact relationship between the two still needs to be resolved through detailed field, petrographic and geochemical work.

3.2.3 Sabieboomrante adamellite gneiss

3.2.3.1 Field relationships

This rock crops out in the southern half of the area, and where it is cut by the Steenbok Shear its outcrop is some 5 km wide narrowing down eastwards to almost a point (see Fig. 3.13, and Annexure 1). De Villiers and Söhnge (1959, p. 107) refer to it as "a well-foliated biotite gneiss".

TABLE 3.21: Geochemical analyses of granodiorites and related rocks
Major element oxides expressed as percentages;
Trace elements in ppm.

	PB622	PB623	PB618	PB616	PB586	2
SiO_2	63.40	64.13	67.30	68.3	61.5	65.6
TiO_2	0.88	0.79	0.76	0.86	1.36	0.57
Al_2O_3	16.05	16.34	14.6	13.97	15.23	15.6
Fe_2O_3*	6.09	5.37	5.74	5.89	7.88	4.7
MnO	-	-	0.1	0.09	0.1	0.08
MgO	2.16	1.88	1.35	1.09	2.3	2.2
CaO	4.66	4.27	2.92	3.28	5.06	4.0
Na_2O	2.95	2.60	2.93	2.36	2.2	3.0
K_2O	3.51	4.37	4.08	3.93	3.85	4.1
P_2O_5	0.29	0.25	0.21	0.23	0.51	0.18
H_2O	-	-				
Total	99.99	100.00	99.99	100.00	99.99	100.03
Rb	128	140	248	187	152	161
Ba	1390	1615	1095	1050	1500	1144
Sr	462	439	188	190	396	435
Th	20	9.1	32.5	12.1	9.8	16
U	-	-		-	-	2.3
Zr	186	196	266	305	364	130
Nb	17	13	22.3	20.2	20.5	8.9
Cr	40	34	5.8	5.9	14.3	28
V	102	87	73	62.8	132	91
Sc	12	12	19.1	16.8	16.3	-
Ni	24	19	15.2	13.2	24.6	18
Co	16	15	18.1	15.5	24.8	13
Pb	30	20	36.5	24.3	16.4	26
Zn	88	73	81.2	83	100	46
Cu	266	75	47.7	15.2	99.4	48
Y	21	30	58.8	72.9	30.7	17

Sample Numbers	Origin	Grid reference Annexure 1
PB 622	Vioolsdrif granodiorite, Kliphoogte pass	L4
PB 623	Vioolsdrif granodiorite, 2km NE of Kliphoogte	M3
PB 618	Sabieboomrante adamellite gneiss	Q9
PB 616	Skurwehoogte granitoid	L12
PB 586	Megacrystic granite	D3
2	Vioolsdrif granodiorite (Sample No. 2, from Reid & Barton, 1983, Table II)	

(Analyst: Dr. D.L. Reid, Department of Geochemistry, University of Cape Town.)

* Total Fe as Fe_2O_3

The rock has a strongly developed foliation enhanced by biotite, and down-dip elongated feldspar augen. The foliation strikes mainly west-northwesterly and dips northwards, although there are local deviations from this pattern. In spite of its tectonised nature the rock protrudes as rounded exfoliation hills typical of granitic weathering. It is greenish-grey and varies between mesocratic and leucocratic varieties, similar to the Vioolsdrif Granodiorite to the north.

Xenolithic rafts occur within the adamellite gneiss although they are rare. The xenoliths comprise a range of country-rock fragments, including cream to pinkish quartzo-feldspathic gneisses, to segmented quartzite units.

Nature of Upper and Lower Contacts

It is very difficult to locate a contact with the adjacent country rock for two main reasons:

- the adamellite gneiss is strongly tectonised at its contact zones;
- the contact zones are mostly covered by scree.

Another factor which obscures a clear-cut contact relationship is the presence of younger granite intrusives, mostly along its northern contact.

However, approximately 3 km west of the old Chabiesies homestead thin sheet-like intercalations of the adamellite gneiss occur in quartzo-feldspathic gneisses and biotite gneisses of the Chabiesies suite. A few hundred metres north of this position grey, recrystallized quartzite occurs as xenolithic units within the adamellite gneiss, outcropping discontinuously for about 3 km and striking north-easterly. The nature of this southern contact is therefore interpreted as intrusive, where competent units of Chabiesies suite country rock have been separated from the main sequence to the south, and now occur as xenolithic fragments within the granitoid (Fig. 3.15).

Another factor consistent with an intrusive interpretation is that north east of the old Chabiesies homestead, the southern adamellite contact transgresses the Chabiesies suite at an acute angle (grid reference T10, Annexure 1). Immediately north of the homestead the same contact is strongly tectonised, obscuring its original intrusive nature.

The northern contact is also markedly tectonized, xenoliths of the Sabieboomrante adamellite gneiss occurring in the Kouefontein granite gneiss.

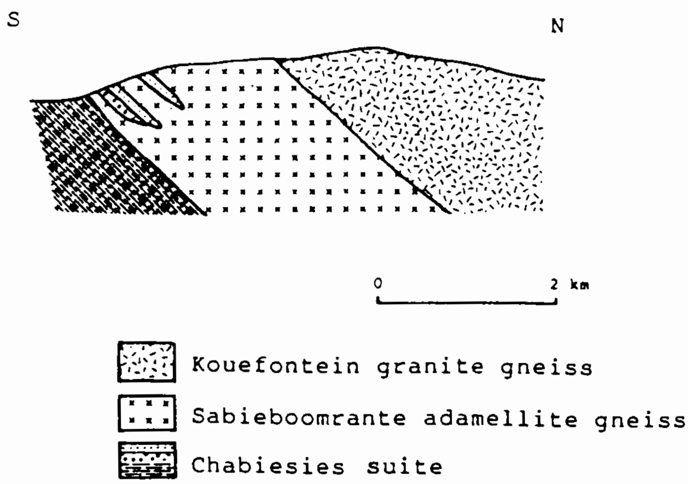


Fig. 3.15 Schematic cross-section showing intrusive nature of the Sabieboomrante granitoid, most notably along its southern contact.

3.2.3.2 Petrography

Modal analyses of this rock and the related Skurwehoogte granitoid west of Steenbok Shear (Fig. 3.13) show that they plot in the granite field on a Streckeisen diagram (cf. Table 3.20; Fig.3.14).

The petrography of this rock-type is summarized in Table 3.22.

This very coarse-grained rock is characterized by prominent augen of quartz, and/or microcline and plagioclase, separated by biotite and accessory epidote, muscovite, magnetite, sphene, apatite, zircon, chlorite and sericite. The fabric is emphasized by a strongly developed $S(N)_2$ biotite and muscovite foliation.

At its southern contact the tectonised fabric is revealed in thin section as
(i) 0,5 mm wide bands of small recrystallized quartz grains outlining an earlier foliation, and
(ii) a conjugate cleavage cutting across (i) at an acute angle (Fig. 3.16).

The cleavage is enhanced by elongate grains of biotite, muscovite, chlorite and sericite. Microcline and plagioclase show signs of fracturing and bending respectively (Fig. 3.16), indicating brittle deformation.

TABLE 3.22 Petrographic Description of Sabieboomrante adamellite gneiss

Texture: Inequigranular.

Quartz: Inequant grains between 0,1 and 3 mm , makes up between 20-30% of the minerals. Larger grains display undulose extinction, smaller grains polygonal and completely recrystallized.

Microcline: Generally large anhedral to subhedral grains, < 3 mm, but grain-size extremely variable. Makes up about 25% of minerals.

Plagioclase: (Oligoclase) forms euhedral to subhedral grains up to 4 mm, making up about one third of the minerals. Occurs in approximately same proportion as K-feldspar, most grains showing advanced saussuritization. Small amount of myrmekitic intergrowth present.

Biotite: Deep olive-green laths < 1 mm long, lepidoblastic texture enhancing fabric of the rock. Slightly replaced by chlorite.

Epidote: Associated mostly with biotite as small anhedral grains, sometimes has an elongated habit oriented along the foliation.

Muscovite: Small euhedral laths < 0,2 mm long, minor component.

Accessories: Magnetite (euhedral and anhedral), apatite, zircon, brown sphene and sericite.

3.2.4 Skurwehoogte adamellite gneiss (related to Sabieboomrante granitoid).

3.2.4.1 Field relationships

This occurs at Skurwehoogte, south-centre of the area, between the Tierkloof and Steenbok Shears (see Fig. 3.13, and Annexure 1). The rock has prominent pink augen of feldspar plunging down-dip and is more melanocratic compared to the Sabieboomrante granitoid to the north-east. The northern contact is similar to that of the Sabieboomrante adamellite gneiss. The southern contact is displaced southwards beyond the area of present investigation, along the Steenbok Shear.

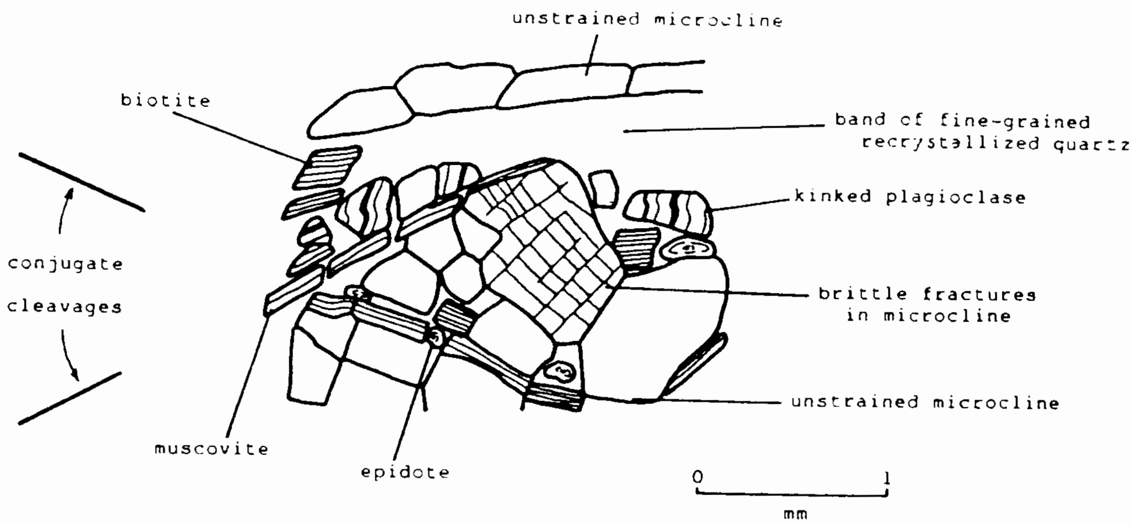


Fig. 3.16 Brittle deformation textures in Sabieboomrante adamellite gneiss. Sketched from a photomicrograph.

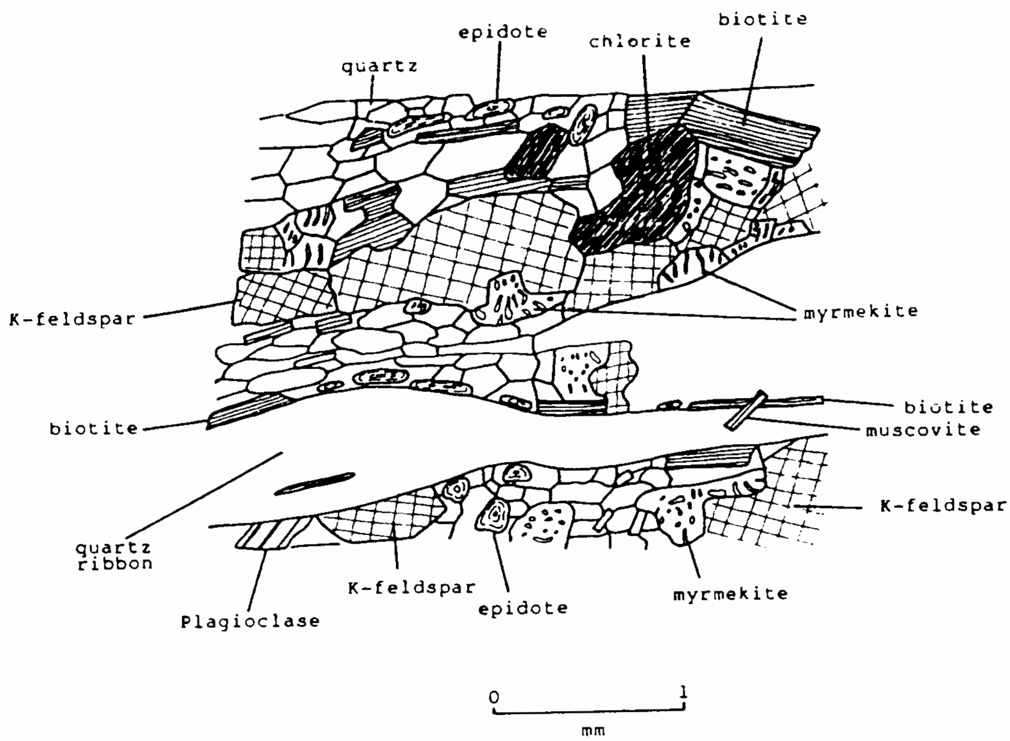


Fig. 3.17 Plastic deformation textures in Skurwehoogte adamelllite gneiss. Drawn from a photomicrograph.

3.2.4.2 Petrography

The Skurwehoogte granitoid has large augen of microcline, saussuritized plagioclase and quartz separated by broad subhedral blades of green biotite, and muscovite. Plagioclase and K-feldspar occur in approximately equal proportions, and biotite makes up about 20% of the minerals. Accessories include epidote, muscovite, magnetite, apatite, zircon and sphene.

Near its northern contact this granitoid displays plastic deformation textures. Biotite and muscovite enclose large feldspar and quartz augen in broad lepidoblastic bands, and, along with smaller recrystallized grains of quartz impart a strongly tectonised fabric. Myrmekite occurs as vermicular intergrowths of quartz and feldspar, and is notably well developed along the margins of the larger feldspar grains.

Immediately west of Tierkloof Shear the Skurwehoogte adamellite gneiss crops out discontinuously whereas farther to the north-west it occurs as elongate northerly oriented xenoliths within the Kouefontein granite gneiss and Kromnek granite (grid reference G4 and G6).

Petrographically they show significant plastic deformation textures (Fig. 3.17). Here quartz grains occur as (i) augen-shaped clusters 4 to 6 mm long by 2 to 3 mm wide, displaying undulose extinction, (ii) as small recrystallized polygonal grains averaging 0,2 mm , arranged in narrow 0,5 mm wide bands separated by the larger grains, (iii) in the form of a broad ribbon structure.

Other minerals include microcline and saussuritized plagioclase in the form of large augen, and olive-green biotite laths 1 mm long, which enclose the augen and emphasize a sigmoidal cleavage. Chlorite replaces biotite along its margins. The abundant development of myrmekite characterizes this rock. Accessories are epidote, muscovite, sphene, iron ore, zircon and apatite.

3.2.4.3 Geochemistry

The Sabieboomrante and Skurwehoogte granitoids have a granodioritic composition but have higher SiO_2 and lower Al_2O_3, MgO and CaO than typical Vioolsdrif Granodiorite (Table 3.21, sample 2).

However, when comparing trace element abundances in the Sabieboomrante granitoid with typical Vioolsdrif granodiorite, Sr is much lower, whereas Zr, Y and Nb are somewhat higher (Table 3.21). Sr and Y plot well outside the field of typical Vioolsdrif Granodiorite on a discriminatory diagram of SiO_2 vs. these elements (D.L. Reid, personal communication, 1986).

3.2.4.4 Other Vioolsdrif Granodiorite-related granitoid bodies

Small sheet-like intrusions similar in field appearance and petrography to the Sabieboomrante adamellite gneiss, occur in the Groothoek and Groenrivier suites (grid references K8, L8, P3, H2, Annexure 1). The outcrop just west of Steenbok Shear (grid reference K8, L8, Annexure 1) contains sizeable biotite-rich xenoliths, but the host rock itself has a gneissic fabric.

Petrographically these rocks contain slightly more quartz (up to 50%), with lesser plagioclase (oligoclase) biotite, epidote, chlorite, apatite and sphene than, e.g., the Sabieboomrante adamellite gneiss.

Broad olive-green laths of biotite outline a sigmoidal cleavage with epidote occurring in abundance, mostly in association with biotite. Apatite forms euhedral six-sided prisms in cross-section, and a minor amount of chlorite replaces biotite.

Thin sheet-like concordant bodies in the Groenrivier suite have a similar appearance and mineral content to the Vioolsdrif Granodiorite. They are greenish-grey, coarse-grained, and composed of microcline, plagioclase, biotite, epidote , apatite and sphene. The strong foliation is outlined by the lepidoblastic arrangement of biotite and epidote.

3.2.4.5 Discussion

The Sabieboomrante and Skurwehoogte granitoids are petrographically and geochemically similar. For correlation purposes if one examines a palinspastic reconstruction to a stage of pre-Pan African tectonism, the Sabieboomrante adamellite gneiss joins up with the Skurwehoogte granitoid, and extends westwards to near the Kromnek Shear (see Chapter 4, Fig. 4.30). The three-dimensional form of these bodies is therefore one of a thick continuous sheet-like intrusion some 30 km long, before it was disrupted by Late Proterozoic/ Early Palaeozoic (Pan-African) tectonism.

The Sabieboomrante and Skurwehoogte granitoids have a similar field appearance and major element geochemistry to Vioolsdrif Granodiorite of the central and northern Richtersveld. However, their stratigraphic position, form of intrusion, and particularly their trace element geochemistry differ quite markedly from typical Vioolsdrif Granodiorite. They are therefore interpreted as a separate suite of rocks, of younger age. Their origin may have a history incorporating some crustal contamination. This interpretation differs from that proposed by De Villiers and Söhnge (1959) who regard these rocks as metamorphosed sediments.

Thin sheet-like granodiorite bodies near the top contact of the supracrustals of the Groenrivier suite are interpreted as apophyses of Vioolsdrif Granodiorite.

3.2.5 Leucogranite

3.2.5.1 Field relationships

This cream to whitish, medium-grained granite occurs at the top of the Groenrivier suite (e.g. grid reference O2, Annexure 1), and is strongly foliated. It forms discontinuous intrusive sheets along strike, attaining a maximum thickness of some 450 m. These sheets have intruded the mafic and ultramafic rocks of the Klipbok complex (Annexure 1) as thin transgressive bodies, but in turn have been tectonised, sometimes into mylonites.

3.2.5.2 Petrography

Modal analyses show this rock has a typical granite composition (Table 3.23, and Fig. 3.14).

The texture is equigranular, and shows some recrystallization of grains. It is composed of quartz, K-feldspar (microcline), plagioclase (oligoclase) and muscovite, with minor epidote, iron ore and garnet.

Near its upper contact this granite is strongly tectonised. Here the fabric is defined by quartz grains of inequant shape flattened parallel to the foliation, and muscovite laths forming discontinuous folia 1 to 2 mm apart throughout the rock. Quartz shows undulose extinction, sutured boundaries between some of the larger grains, and a mortar texture. Subhedral microcline grains up to 5 x 2 mm in size make up 15 to 20% of the minerals present.

Subhedral to anhedral, variably saussuritized plagioclase grains constitute 30% of the minerals. This secondary alteration is taking place from the inside, leaving a clear outer rim.

3.2.5.3 Geochemistry

A geochemical analysis reveals a high SiO_2 content of 75,11%, with low abundances of the trace elements Zr, Sr, Ba and Y (Table 3.24, Sample PB 620). It is a leucocratic alkali granite and differs from all other granites in the area, particularly in its trace element content. In comparison to average Richtersveld Suite granites Zr is particularly low, viz. 32 ppm vs 100 ppm (Table 3.24, Sample 4). This could be the result of magmatic differentiation, but may also reflect evidence of weathering (D.L. Reid, personal communication, 1986). Rb in this rock is high compared to the

TABLE 3.23: Modal composition of Kouefontein granite gneiss○, Leucogranite⌖ and Kromnek granite⊙.
Analyses carried out with a Swift Point Counter.

Mineral	PB153○	PB181○	PB187○	PB200⌖	PB201⌖	KROM1⊙	PB606A⊙	PB617⊙
Quartz	25	32	25	39	34	38	30	35
Alkali Feldspar	61	37	45	31	37	40	45	40
Plagioclase	7	21	16	16	17	12	10	11
Biotite	–	2	–	–	–	–	2	2
Muscovite	2	8	12	13	12	5	4	6
Chlorite	–	–	2	–	–	4	5	2
Epidote	–	–	–	–	–	1	3	2
Sphene	–	–	–	–	–	tr	1	–
Iron ore	tr	tr	–	–	–	–	–	0.5
Apatite	–	–	–	–	–	tr	–	0.5
Sericite	tr	tr	–	–	–	–	–	–
Garnet	5	–	–	1	tr	–	–	–
Calcite	–	–	–	–	–	–	–	1
TOTAL	100%	100%	100%	100%	100%	100%	100%	100%

TABLE 3.24: Geochemical analyses of granitic rocks. Major element oxides are expressed in percentages; Trace elements in ppm.

	PB575	PB619	PB617	PB620	4
SiO_2	71.90	72.10	70.46	75.11	75.60
TiO_2	0.20	0.19	0.41	0.06	0.20
Al_2O_3	14.91	14.96	15.09	14.03	13.00
Fe_2O_3 *	1.87	1.99	2.82	1.08	1.20
MnO	-	-	-	0.09	0.03
MgO	0.29	0.34	0.59	0.14	0.29
CaO	1.12	1.25	1.26	0.35	0.83
Na_2O	3.36	3.44	3.59	3.33	3.50
K_2O	6.25	5.60	5.63	5.58	5.30
P_2O_5	0.10	0.12	0.15	0.23	0.04
TOTAL	100.21	99.97	99.68	100.56	99.99
Rb	293	306	218	353	246
Ba	627	1190	1950	247	521
Sr	137	218	276	61.2	101
Th	66.6	41.4	30.4	-	28
U	-	-	-	-	4.9
Zr	202	202	364	32.1	124
Nb	10.1	15.1	14.2	22.1	17
Cr	-	3.6	2.7	6.4	2.6
V	-	10.4	29.4	2.2	10
Sc	5.4	3	2.5	6.1	-
Ni	1.6	2.1	4.7	1.6	3
Co	3.6	4.7	6.8	-	3.7
Pb	42.4	40.4	19.3	29.1	35
Zn	35.3	39.3	49.7	11	12
Cu	154	46.4	155	11	18
Y	27.2	9.6	5.2	23.1	23

Sample Numbers	Origin	Grid reference, Annexure 1
PB 575	Kromnek granite	G6
PB 619	Donkiehel granite	K8
PB 617	Biotite granite, Skurwehoogte	L12
PB 620	Pretectonic leucogranite	O2
4	Leucogranite (from Reid and Barton, 1983, Table II).	

(Analyst Dr. D.L. Reid, Department of Geochemistry, University of Cape Town).

* Total Fe as Fe_2O_3

Richtersveld granites, viz. 353 ppm vs 250 ppm. From its trace element content this granite does not resemble average Richtersveld Suite granites, but is similar to typical leucogranites from the northern Richtersveld. It has a high Rb/Zr ratio similar to silica-rich muscovite leucogranites associated with syntectonic deformation, in collision belts such as the Himalaya of Asia and the Hercynian of Europe (Harris et al., 1986).

3.2.6 Dabbieputs Granite Gneiss

3.2.6.1 Field relationships

This buff coloured, coarse-grained granite occurs as narrow, elongate sheets mainly along the contact between the Groothoek and Ratelfontein suites. In stratigraphically higher locations within the Groothoek suite this granite is particularly coarse-grained, and becomes a pegmatitic granite gneiss.

3.2.6.2 Petrography

The rock is very coarse-grained, has an equigranular texture, and a foliation defined by the faint lepidoblastic alignment of muscovite and chlorite. Strain effects are notable in quartz and plagioclase grains. Inequant quartz averaging 1 mm , has undulose extinction within grains, and suturing between them. Smaller polygonal recrystallized grains which surround larger strained ones form a mortar texture. Large saussuritized oligoclase averages 1,5 mm, and make up about 40% of the minerals present. Minor microcline, forming subhedral to anhedral grains up to 2 mm, as well as biotite, muscovite, chlorite and sericite are present.

Muscovite has two habits:
- slender laths aligned along the foliation;
- broad irregular grains notably kinked, displaying undulose extinction, and sometimes poikiloblastically enclosing quartz.

3.2.6.3 Discussion

West of Kromnek Shear the granite has the same field appearance as that east of the Steenbok Shear, and is correlated with the latter because of its similar stratigraphic position just above the Ratelfontein suite. This correlation may prove to be incorrect as the granite here resembles the Kouefontein granite gneiss to its south. Further detailed work, however, is needed for verification.

3.2.7 Tweeriviere Granite Gneiss

3.2.7.1 Field relationships

This granitic rock is located just west of Kromnek Shear where it intrudes Vioolsdrif Granodiorite, segmenting the latter into west-northwesterly striking xenolithic rafts (Fig. 3.13, and Annexure 1). It separates Vioolsdrif Granodiorite from the metavolcanics of the Windvlakte suite, but in turn is very strongly tectonised along a broad thrust zone at the base of the metavolcanics (grid reference F3 and G3, Annexure 1).

The rock is a buff to pale pinkish coloured, medium to coarse-grained granite gneiss, with a well developed $S(N)_2$ foliation. It is composed predominantly of quartz, feldspar and minor muscovite.

3.2.7.2 Petrography

Small anhedral microcline grains are poikilitically enclosed by very large plagioclase (oligoclase) grains, while muscovite forms subhedral laths parallel to the predominant foliation within the rock. Some quartz grains show intragranular deformation, but a minor proportion of recrystallized grains is also present.

3.2.7.3 Discussion

Although the Tweeriviere granite is younger than Vioolsdrif Granodiorite its relationship to the megacrystic granite is not clear, as both have been strongly tectonised along their contact zones. As the megacrystic granite is intimately related to Vioolsdrif Granodiorite and interpreted as younger than the latter (see section 3.5.2), it is quite probable that the Tweeriviere granite is younger than the megacrystic granite. This suggestion needs to be confirmed, as information at present is not adequate to prove this interpretation.

3.2.8 Granite gneiss associated with Lower Unit, Windvlakte metavolcanics

A fine- to medium-grained granite gneiss occurs intercalated as minor sheets in the Lower Unit, Windvlakte suite. The granite gneiss is composed predominantly of plagioclase, microcline and quartz, and has an inequigranular texture. Accessory minerals include muscovite, chlorite, epidote and iron ore. The plagioclase grains are heavily saussuritized, and quartz displays undulose extinction.

These granitic sheets occur in much the same stratigraphic position as the Tweeriviere granite gneiss, and can probably be correlated with the latter.

3.2.9 Granite gneisses associated with the Chabiesies suite

Banded biotite gneiss

This medium-grained rock is pale green to grey in colour and has a distinctly banded field appearance due to the presence of numerous thin quartzo-feldspathic leucosomes within the biotite quartz feldspar host. The leucosomes were folded during a phase of ductile deformation, with subsequent intrusion of quartzo-feldspathic veins parallel to the axial-planar foliation. This rock could possibly be correlated with the Brandewynsbank or Noenoemaasberg gneiss, from descriptive accounts in the literature (Van Aswegen, 1983; 1988; Van der Merwe, 1986), although this still has to be proved.

Beige nodular quartz-feldspar gneiss

Prominent in the far south-east corner of the area, this foliated rock is composed predominantly of quartz and feldspar with minor muscovite and sillimanite. Occasional nodules of quartz and sillimanite up to 3 cm, are sporadically developed. A prominent pervasive cleavage, similar to that found in the previously described rocks, was formed during the last retrograde metamorphic event here.

This rock has an inequigranular texture composed of medium-grained quartz, plagioclase and microcline, with occasional muscovite porphyroblasts. Smaller muscovite laths approximately 1 mm or less in size are also present. In some thin sections chlorite has replaced muscovite laths to a marked degree. As in most of the pelitic rock-types in this area sericite has replaced sillimanite and feldspar, forming a large proportion of the minerals present. It notably occurs along the prominent foliation.

This rock is a correlate of the Noenoemaasberg gneiss of the Gladkop Suite (W. Basson, personal communication, 1987). From descriptive accounts by Van Aswegen (1983; 1988) similar rocks in the Namaqua geotraverse, the nodular quartz-feldspar gneiss may be a correlate of the Kinderlê gneiss. The latter proposal, however, needs verification.

In order to solve correlation problems in this area a regional mapping programme should be initiated to include the area of the Namaqua geotraverse and continue westward to the easternmost outcrops of the Gariep Group. Isotopic dating of key rocks in this region should be carried out simultaneously in order to elucidate the chronological history of the region.

3.2.10 Kouefontein granite gneiss

3.2.10.1 Introduction

Referred to as "muscovitised granite" (De Villiers and Söhnge, 1959), "aplogranitic gneiss" (Joubert, 1971) and the "Pink-gneiss Zone" (Ritter, 1980, p. 68), these rocks outcrop over a broad area. East of 17°30′ they are called "muscovite gneiss" by Theart (1980), and leucogranite by Blignault et al. (1983). As the current investigation traverses the Bushmanland and Richtersveld Subprovinces, a better perspective of the characteristics, distribution and possible mode of emplacement of the granite gneiss has been gained.

The Kouefontein granite gneiss comprises mainly a leucocratic granite gneiss and occupies the largest outcrop area of all rock-types encountered in the present survey. It forms the high ground which makes up the north-westerly trending watershed. Its most continuous outcrop is east of Steenbok Shear from where its form tapers eastwards, and eventually dies out some 25 km beyond the boundaries of the present area. West of this shear its outcrop pattern is markedly offset along the Kromnek and Tierkloof Shears (cf. Fig. 2.5, and Annexure 1).

The broadest outcrop area is in the extreme south-west where it is partially covered by the younger Stinkfontein Formation. Its northern contact is remarkably concordant with the Ratelfontein suite, but does transgress this unit as small sheets. The southern contact is concordant east of Steenbok Shear, but becomes transgressive westwards. About 4 km north of the old Chabiesies homestead, xenoliths of Sabieboomrante adamellite gneiss occur within the Kouefontein granite gneiss, indicating their relative age. The granite gneiss is strongly tectonised along sections of the southern contact where it acquires an augen texture in places.

Xenolithic rafts are an important characteristic. In particular these comprise Chabiesies suite metasediments, occurring in the east. There are also a variety of other xenoliths including quartzite, amphibolite and one block of Vioolsdrif Granodiorite measuring some 60 m long (grid reference P5, Annexure 1).

Another characteristic is the presence of relatively abundant, randomly oriented large muscovite porphyroblasts. De Villiers and Söhnge (1959, p. 112) ascribe the growth of this muscovite to K-solutions introduced during the hydrothermal stage of pegmatite development in this region.

3.2.10.2 Lithologies

These rocks are relatively homogeneous, but where a northward dipping $S(N)_2$ foliation transects the pluton alternating foliated and unfoliated bands occur. The foliated zones contain biotite in addition to muscovite, which categorizes these rocks into two predominant types.

(i) muscovite granite-gneiss

This is a whitish to beige, coarse-grained rock, with large flakes of muscovite. Massive varieties tend to occur predominantly towards the top contact. The rock has an inequigranular texture. Quartz occurs as large anhedral grains displaying strain shadows and sutures between them. Other essential minerals include subhedral plagioclase, microcline, and microcline-microperthite. Euhedral to rounded garnet grains, generally 1 mm in diameter, and broad irregular laths of muscovite with random orientation, occur as late-phase porphyroblasts. Minor oval-shaped epidote grains and stringers of magnetite are also present. Sericite is scarce in this rock-type, but does replace muscovite, and also occurs within cracks in garnet.

(ii) biotite granite-gneiss

This variety of granite gneiss differs from the previous type in that it possesses a marked foliation. Its fabric shows the development of biotite along the foliation accompanying a slight diminution in grain size. The rock displays a parallel alignment of large subhedral microcline plagioclase and quartz grains, with sutured boundaries between them. Folia made up of muscovite and biotite laths and spaced 3 mm apart on average, separate the large grains.

Modal analyses of each of the types plot in the granite field on a Streckeisen diagram (Table 3.23, and Fig. 3.14).

3.2.10.3 Discussion

There are some important aspects regarding the Kouefontein granite gneiss which need to be noted. Firstly, there can be no doubt that these rocks are intrusive. Evidence provided by xenoliths testifies to this, although the granitic rocks themselves have subsequently been deformed and now posses a tectonised fabric. These rocks are younger than the Sabieboomrante granitoid, but older than the Kromnek Suite of K-granites (Spektakel Suite age?) and pegmatites. Their age can therefore be bracketed between about 1800 to 1100 Ma - the latter constraint being the Namaqua event. Secondly, the structural position of the Kouefontein granite gneiss is significant with regard to the following aspects:
- it separates lithologies of the Ratelfontein suite in the north from those of the Chabiesies suite in the south. At its northern contact the Ratelfontein suite clearly has an approximate westerly strike whereas at its southern contact the Chabiesies suite has a

northerly orientation west of Tierkloof Shear (e.g. Grid Reference I 13, Annexure 1). East of Steenbok Shear the Chabiesies suite has only a slight divergence south of the Sabieboomrante adamellite gneiss, compared to lithologies to the north. It would appear that the Kouefontein granite gneiss, along with the Sabieboomrante adamellite gneiss, has exploited a zone of either lithological or structural unconformity here. The pluton is envisaged as intruding along this discordant zone, uplifting the roof and incorporating xenolithic blocks, and causing the base of the pluton to acquire a tectonic fabric. The latter feature is a significant characteristic of the Kouefontein granite gneiss, indicating that at least part of the pluton is probably syn-tectonic;
- the outcrop has a wedge shape, being widest in the west and tapering eastwards to a point (Fig. 3.13 and Annexure 1). I speculate that the source of the Kouefontein granite was to the west.

3.2.11 Granite gneisses occurring south of the Kouefontein granite gneiss

Large discordant leucocratic granite gneisses occurring well south of the Kouefontein granite gneiss sheet, but of similar field appearance, crop out in the south-eastern extremity of the present area. Investigation into the petrography and geochemistry of these bodies is warranted in order to correlate them with plutons east of 17°30′. They are tentatively correlated with the Eyams granite, based on continuity of the latter granite from Theart′s (1980) mapping westwards to join up with leucocratic granite gneisses to the south of the present area.

3.2.12 K-granites

3.2.12.1 Field relationships

These granitic bodies occur as stocks and sheets of variable size, intruded into the "Transitional Zone" and the Bushmanland Subprovince. Only one small circular stock transects the Vioolsdrif Granodiorite near Kliphoogte (grid reference K4, Annexure 1). The largest pluton, a pinkish, medium- to coarse-grained granite measuring 2 km in diameter, occurs at Kromnek in the western portion of the area (grid reference H6, Annexure 1). The remaining intrusions are very much smaller circular to oval-shaped bodies, although concordant sheets are also common. They have a smooth exfoliation or sometimes rounded boulder form of weathering. Their variability in grain-size and colour depends on the type of K-feldspar and amount of biotite present.

A common characteristic includes xenolithic blocks of biotite quartz-feldspar gneisses, especially within the Kromnek granite (De Villiers and Söhnge 1959, p. 76). This pluton post-dates $S(N)_1$

and $S(N)_2$, but predates the pegmatites, giving it an age slightly greater than 1000 Ma (Ritter, 1980).

3.2.12.2 Petrography

Modal analyses show these rocks are alkali-rich granites (Table 3.23; and Fig. 3.14).

Essential minerals comprise quartz, microcline, microcline-microperthite and plagioclase, with accessory muscovite, biotite, epidote, sphene, apatite, chlorite and calcite.

The rocks have an inequigranular texture where large quartz grains often occur in clusters and show undulose extinction, whereas small grains have a polygonal shape. Saussuritized plagioclase has small muscovite lath inclusions displaying a nematoblastic growth. Microcline and microcline-microperthite can be up to 6 mm long. Myrmekite is an important accessory. Biotite occurs as small subhedral dark brown laths, slightly replaced by chlorite. Epidote and calcite form irregular-shaped grains, occurring more abundantly along fractures. Sphene, iron-ore and apatite are present in minor amounts and are preferentially associated with the phyllosilicates and calcite. Muscovite forms short broad laths.

3.2.12.3 Geochemistry

Geochemical analyses of three separate intrusions, viz., Kromnek (grid reference G6, Annexure 1), Donkiehel (grid reference K8, Annexure 1) and a biotite granite near Skurwehoogte (grid reference L12, Annexure 1) are presented in Table 3.24. There is very little variation in major element and trace element abundances between the three separate bodies.

These granites are, however, different to granites typical of the Richtersveld Suite from the area farther north, when comparing discriminatory trace elements Zr, Nb, Y and Sr. (D.L. Reid, personal communication, 1986). The Kromnek (PB 474), Donkiehel (PB 619) and Skurwehoogte biotite granite (PB 617) show higher Zr, Nb and Y, and lower Sr than typical Richtersveld Suite rocks (see Table 3.24). They do not necessarily therefore compare with the Richtersveld Suite granites, as they are of a slightly different geochemical type, and may in fact define a geochemical gradient not previously recognized (D.L. Reid, personal communication, 1986).

3.2.12.4 Discussion

The Kromnek granite is cut by north-northeast trending felsic (bostonite) dykes which are of syenitic composition, and comagmatic with plutonic members of the Richtersveld Suite (Middlemost, 1963). These dykes can therefore probably be assigned a 920 Ma age similar to that of the Richtersveld Suite granites (Allsopp et al., 1979).

The Kromnek granite is the only one where felsic dykes transgress plutons of this suite, and in view of what was said above, it would appear that the K-granites define a suite of intrusives older than 920 Ma. These observations corroborate the age relation ships proposed by Ritter (1980). In this case the term Kromnek Suite should be applied for the younger K-granites in this area.

Plutons west of Steenbok Shear show recrystallization inferred to be related to the Pan-African metamorphic event. The Kromnek granite, for example, has a recrystallized texture in which large inequant microcline-microperthite grains occur in a recrystallized groundmass of quartz, plagioclase and biotite. Plagioclase (oligoclase) forms smallish euhedral to subhedral grains, whereas quartz displays a typical granoblastic-polygonal texture, with straight boundaries. Biotite occurs in randomly oriented irregular to subhedral laths, and is slightly replaced by chlorite. Sericite is present between other grains and in cracks in the large feldspar grains. Minor irregularly-shaped iron-ore and epidote grains are also present.

The spatial distribution and form of intrusion of the K-granites are remarkably similar to the Wyepoort granite mapped by Theart (1980) and Blignault et al. (1983) east of the present area (cf. Fig. 2.5). This relatively undeformed granite suite occupies a significant spatial and temporal position with regard to the evolution of the zone separating the Richtersveld and Bushmanland Subprovinces because it occurs both in the hangingwall and footwall of the Groothoek Thrust zone, thus postdating the thrusting.

3.2.13 Amphibolite

Amphibolites of the Klipbok complex occur east of the Kouefontein Shear, and are associated with the Groenrivier suite. In the extreme north-east of the area, the Klipbok mafic/ultramafic complex crops out discontinuously for at least 7 km along a north-westerly trending linear zone at the top of this unit (see Plate 8, and Annexure 2). The maximum width of the zone is just less than 1 km.

3.2.13.1 Lithological relationships

Amphibolite forms the main body of the Klipbok complex and shows some variable characteristics, especially with regard to grain size and a gross overall layering. In the west outcrops are black, and where large plagioclase and amphibole minerals are more abundant, the rock is spotted. In the east both coarse and medium-grained varieties of amphibolite overly a cordierite schist unit, and appear strongly tectonised (grid reference R4, Annexure 1). A maximum thickness for the western-most amphibolite unit is 180 m.

The main minerals are hornblende and plagioclase (andesine), the latter showing replacement by sericite. Accessories include quartz, epidote, muscovite and clinozoisite. The rock has a variable inequigranular to equigranular texture. A crude layering throughout the amphibolite is due to:

- grain size variation upwards, and locally in the sequence;
- a higher concentration of certain minerals, e.g. hornblende and iron ore, within particular zones.

Five zones make up the layering in the amphibolite (see Fig.3.18).

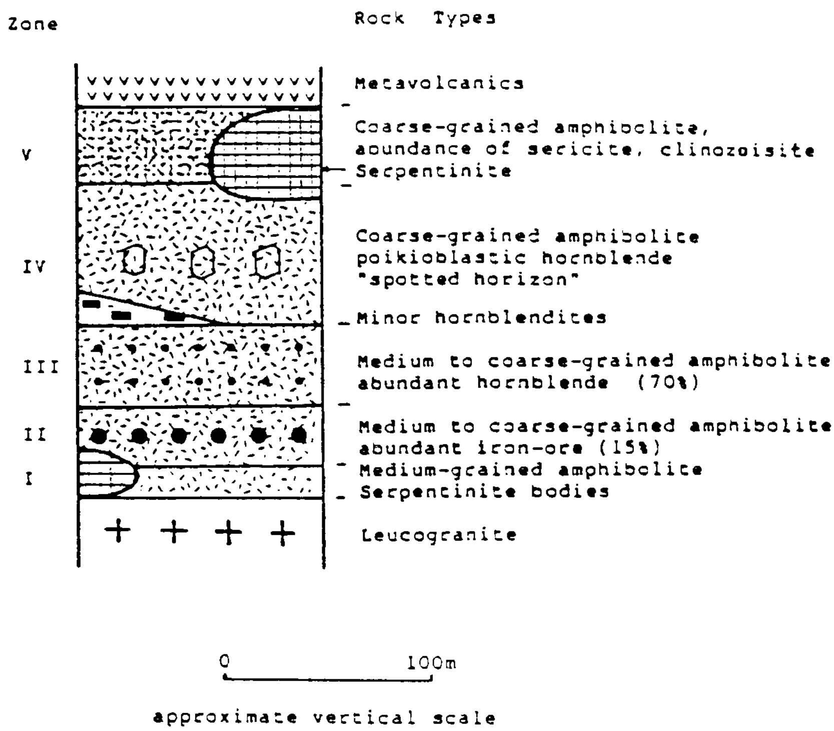


Fig. 3.18 Schematic section, Klipbok mafic/ultramafic sequence.

A petrographic description of the amphibolites of the Klipbok complex is given in Table 3.25.

TABLE 3.25 Petrography of amphibolites, Klipbok complex	
Zone	**Description**
V (top)	Coarse-grained amphibolite shows extremely advanced stages of replacement. Sericite completely replaces plagioclase, whereas hornblende and muscovite have been affected to a lesser degree. Clinozoisite fairly abundant;along with chlorite has undulose extinction.
IV	Central and upper portions of amphibolite made up of a coarse-grained rock with a notably spotted appearance. Plagioclase and hornblende occur in approximately equal proportions and are of about equal size, sometimes forming decussate clusters. Where lesser plagioclase present then thin banding giving rise to a crude layering. Accessories include epidote, muscovite, chlorite and clinozoisite. The latter mineral has a characteristic blue-grey colour, contains abundant small inclusions, and forms broad blades up to 1,8 mm long by 0,5 mm wide. Plagioclase, (labradorite) shows Carlsbad, Albite and Pericline twinning. Blue-green hornblende poikiloblastically encloses small recrystallized quartz grains.
III	Zone contains only ±5% iron ore, yet retains high proportion of hornblende. Plagioclase here has been altered to epidote and muscovite. Magnetite forms euhedral crystals whereas ilmenite is anhedral.
II	A large proportion of hornblende (70%) occurs as porphyroblasts (<3 mm), showing exsolution of opaque minerals parallel to the prismatic crystallographic plane. Rock is more iron-rich than other varieties, containing about 15% iron ore. Plagioclase is much smaller than in Zone 1.Epidote occurs abundantly as granular small and large blebs. A minor amount of quartz is poikiloblastically enclosed by hornblende.
I base)	Medium-grained, composed of plagioclase, blue-green hornblende, epidote, clinozoisite, muscovite and chlorite. Hornblende makes up ±55% and plagioclase (labradorite) ±30% of the rock. Epidote is abundant although not in contact with plagioclase. Muscovite and clinozoisite are accessory.

The eastern-most amphibolite, which occurs in association with a cordierite schist unit, comprises both coarse-grained and fine-grained varieties.

(i) Coarse-grained variety

The rock is extremely inequant, composed of plagioclase (labradorite) and blue-green hornblende, with accessory quartz, epidote, iron ore and sphene. Plagioclase and quartz exhibit a granoblastic-polygonal texture whereas hornblende forms small euhedral prisms and prominent irregular poikiloblasts 5 mm in size.

The amphibolite is strongly tectonised in places, the fabric expressed as folia of elongated 1 to 2 mm long subhedral hornblende laths. The folia are separated by inequant plagioclase grains less than 1 mm in size. Lime-yellow anhedral epidote grains, forming elongate clusters up to 5 mm long, along with hornblende, are aligned parallel to the foliation. Near the top of this unit the rock shows decussate clusters of randomly oriented, mostly euhedral, blue-green hornblende grains, separated by a felted mass of sericite. Plagioclase here is completely saussuritized. Clinozoisite is accessory.

(ii) Fine-grained variety

This equigranular, fine-grained (average 0,1 mm) rock occurs as narrow bands in the coarse-grained type, plagioclase (andesine) and quartz forming a granoblastic-polygonal texture. Elongate and fibrous muscovite laths up to 1 mm long are distinctly bent. Accessories include small anhedral epidote and chlorite grains.

Sericite is present to varying degrees, even in one thin section. It can occur as very small laths between and replacing the polygonal plagioclase grains, but elsewhere it has replaced all the original minerals completely.

3.2.13.2 Geochemistry

The coarse-grained "spotted" amphibolite from the Klipbok complex has a major element composition which is not directly comparable with average tholeiitic or calc-alkali mafic rock types (Table 3.26, samples PB 502; 1, 2, 3 and 4). Al_2O_3, CaO and MgO in the amphibolite are slightly higher and Fe_2O_3, Na_2O_3 lower than in these rock types. This can probably be attributed to the high proportion of plagioclase and epidote in amphibolites from the Klipbok complex.

TABLE 3.26: Geochemical analysis of amphibolite from the Klipbok complex compared to analyses of mafic rocks from the literature.
Major element oxides expressed as percentages; Trace elements as ppm.

	PB502 #	1	2	3	4
SiO_2	49.96	49.8	51.1	50.3	50.50
TiO_2	0.31	1.5	0.83	2.2	0.80
Al_2O_3	18.20	16.0	16.1	14.3	15.60
Fe_2O_3	7.11*	10.0	11.8	13.5	0.90
FeO	-	-	-	-	9.30
MnO	0.11	-	-	-	0.20
MgO	9.88	7.5	5.1	5.9	8.50
CaO	12.42	11.2	10.8	9.7	10.80
Na_2O	1.23	2.75	1.96	2.5	2.00
K_2O	0.75	0.14	0.40	0.66	0.60
P_2O_5	0.04	-	-	-	0.10
H_2O	-	-	-	-	0.80
TOTAL	100.01	98.89	98.09	99.06	100.10
Rb	74.6	1	5	31	
Ba	127	11	50	170	
Sr	374	135	225	350	
Th	6.1	0.18	0.5	1.5	
U	-	0.10	0.15	0.4	
Zr	28.6	85	60	200	
Nb	-	-	-	-	
Cr	154	300	50	160	
V	118	-	-	-	
Sc	37.8	-	-	-	
Ni	193	100	25	85	
Co	48	32	20	38	
Pb	7.7	-	-	-	
Zn	47	-	-	-	
Cu	69.5	-	-	-	
Y	9.8	-	-	-	

Sample Numbers	Origin	Grid reference, Annexure 1
PB 502	Amphibolite (metagabbro), Klipbok complex.	02
1	Average low-K rise tholeiite (from Condie, 1976, p. 148)	
2	Average low-K arc tholeiite (from Condie, 1976, p. 148)	
3	Average continental rift tholeiite (from Condie, 1976, p. 148)	
4	Average Kokstad type Karoo Dolerite (from Walker & Poldervaart, 1949, p. 666)	

(# Analyst: Dr. D.L. Reid, Department of Geochemistry, University of Cape Town.)

* Total Fe as Fe_2O_3

A comparison of some of the less mobile trace elements, e.g. Co, Cr, Zr, Ti from basalts of different tectonic settings reveals that there is no consistent correlation between Klipbok complex amphibolites and possible equivalent rock types formed in either divergent or convergent tectonic environments (Table 3.26, samples 1 and 2).

3.2.13.3 Discussion

Significant features of the Klipbok complex amphibolites include the following:
- layering, defined by subtle variations in proportions of minerals present rather than change in rock-type. The iron-rich layer near the base, and the coarse-grained plagioclase-rich variety higher up in the unit suggest that an initial cumulus settling and differentiation probably took place within this gabbroic unit;
- strongly developed cleavage/foliation, especially within the south-easternmost occurrence, suggests deformation related to the Namaqua event;
- greater abundance of sericite at the top of the unit than elsewhere, which suggests proximity to a zone of retrograde metamorphism.

Metamorphism during the Namaqua event is likely to have caused greater mobility of certain elements compared to others. Interpretations of geochemical results regarding the origin of the Klipbok complex rocks is therefore inconclusive. A similar problem is encountered in mafic rocks of calc-alkali affinity in the Sabzevar ophiolite range, north-east Iran (Lensch, 1979). Deuteric alteration and low grade metamorphism of the mafic rocks here has resulted in a change in major element geochemistry. This makes it difficult to interpret the original environment of formation of the mafic rocks. Even though conclusions cannot be satisfactorily reached using present geochemical data there may be some analogy between amphibolites of the Klipbok complex and mafic rocks formed in an island arc environment. Mafic rocks of mature island arcs are described by Burns (1985) and Leake (1989) as having a high Al_2O_3 and CaO, and low TiO_2 geochemical signature, similar to those of the Klipbok complex.

3.2.14 Serpentinites

The serpentinites are associated with two distinct zones, viz. the Klipbok complex in the Groenrivier suite, and the Kouefontein granite gneiss.

3.2.14.1 Field relationships

3.2.14.1.1 The Klipbok complex occurrences

Here the serpentinites occur as (i) small oval to sheet-like entities strung out for at least 4 km along a north-westerly trending lineament at the top of the Groenrivier suite, and/or (ii) as irregular lenses intimately associated with the amphibolitic rocks, especially at the top, but also at the base of the unit (see Annexure 2). One isolated pod occurs within Vioolsdrif Granodiorite just north of this zone. The largest of these bodies is of the order of 200 m by 30 m on outcrop.

When viewed from a distance a northward-dipping layering is evident in some of these rocks, although this is not that obvious on a mesoscale. The ultramafics are easily recognized on aerial photographs due to their dark tone, and in the field they have a contrasting blackish to reddish-brown colour, and characteristic knobbly, rough-weathering surface. Most of them protrude as positive topographic features, but at their contacts they are highly weathered, thereby obscuring the all important relationship with the country rock.

A fresh specimen of serpentinite is dark greyish-green to black, and is very coarse grained. Serpentine (antigorite) occurs as a grey felted mass of small grains replacing olivine, and makes up about 60% of the minerals. Original olivine grains are outlined by the irregular distribution of magnetite and chlorite along characteristic remnant cracks. Magnesium chlorite occurs as light grey elongate twinned grains sometimes in broad laths up to 2,0 mm in size. Grains are mostly bent and strained, and to a large extent replaced by sericite. Iron ore is intimately associated with chlorite, occurring as exsolution grains within the chlorite. Tremolite forms irregularly-shaped grains up to 5 mm in size. Most show exsolution lamellae of opaque ore parallel to the prismatic faces. The mineral also occurs as long slender needles, generally 1 mm long. Magnetite occurs as euhedral, but mostly as anhedral grains making up approximately 25% of the minerals. A small amount of talc is present.

3.2.14.1.2 Occurrences within the Kouefontein granite gneiss

Three small bodies of serpentinite (smallest 2 m diameter), similar to the Klipbok complex occurrences, occur in the Kouefontein granite gneiss (grid references N5, P5 and T9, Annexure 1). One of these bodies (adjacent to the Kouefontein Shear) has the same petrographic features as those of the Klipbok complex, but in addition shows advanced replacement by sericite.

3.2.14.2 Geochemistry

A serpentinite from the Klipbok complex and one from the Kouefontein granite gneiss have similar major element abundances (Table 3.27). SiO_2 and MgO within the latter are slightly higher, whereas Al_2O_3 and CaO are slightly lower, probably reflecting a late stage metasomatic alteration. Trace element abundances reveal anomalous concentrations of Cr and Ni in the Klipbok sample (PB 504), whereas the sample from the serpentinite within the Kouefontein granite gneiss (PB 495) contains higher proportions of Rb, Sr, Zr, Cr, Pb, Cu and Y. There appears to be a strong depletion in Ni, V and Ba. In spite of these geochemical differences the two rocks probably have a common origin, but may have experienced slight metamorphic alteration.

Comparing the geochemistry of the Klipbok ultramafics with those from an area near Rooiberg 2 some 20 km to the north of the present area, the former have a higher SiO_2 and CaO content than the latter (see Table 3.27, Sample 1). This can probably be ascribed to alteration associated with zones of high shear strain in the present area. The serpentinites from the Rooiberg 2 area probably represent crystallization differentiation products of a mafic magma (Middlemost, 1963, p. 67).

One sample of ultramafic rock from the Klipbok complex (PB 504) has low TiO_2 (0,19%) and Na_2O (0,03%) and high Al_2O_3 (6,04%), similar to ultramafic rocks of the Dras area, Ladakh, India (Prasad, et al., 1979). The latter are intrusive into basaltic rocks and siliceous/tuffaceous sediments.

3.2.15 Hornblendites

3.2.15.1 Lithological relationships

These occur as small circular to oval sheet-like bodies intimately associated with serpentinites and amphibolites of the Klipbok complex, but sometimes as entities, e.g. 4 km south-east of the Klipbok complex. They are black to dark green, very coarse-grained rocks, composed almost entirely of pale blue-green hornblende, with negligible accessory quartz, iron ore, epidote and talc. The grains are of extremely variable size, giving the rock a decussate texture. Hornblende here has the composition $(Ca, Na)_{2.26}$ $(MgFeAl)_{5.15}$ $(Si,Al)_8$ $(O)_{22}$ $(OH)_2$ (determined by X-ray diffraction and matching 2θ peaks against those determined for hornblende in the Data Book; Selected Powder Diffraction Data for Minerals (1974); see Appendix A -1.4).

In the area around Rooiberg 2 hornblendites (90% amphibole, with minor chlorite, epidote, plagioclase and white mica) are more abundant than serpentinites (60% serpentine with lesser amounts of

TABLE 3.27: Geochemical analyses of ultramafic rocks.
Major element oxides expressed as percentages;
Trace elements as ppm.

	PB504 #	PB495 #	1
SiO_2	43.85	47.92	38.50
TiO_2	0.19	0.17	0.29
Al_2O_3	6.04	2.32	2.76
Fe_2O_3	12.73*	11.24*	7.96
FeO	-	-	5.47
MnO	0.13	0.13	0.20
MgO	33.56	37.23	32.68
CaO	3.41	0.95	1.57
Na_2O	0.03	0.01	0.03
K_2O	0.03	-	0.16
P_2O_5	0.02	0.02	0.07
H_2O	-	-	7.96
TOTAL	99.99	99.99	100.54
Rb	-	325	
Ba	59.1	19.2	
Sr	30.9	150	
Th	-	71.7	
U	-	4.4	
Zr	14.3	217	
Nb	-	12.3	
Cr	3610	4685	
V	83.8	34.7	
Sc	17.5	12.4	
Ni	2265	-	
Co	148	141	
Pb	-	48.2	
Zn	92.6	40.4	
Cu	28.7	174	
Y	4.7	29.1	

Sample Numbers	Origin	Grid reference, Annexure 1
PB 504	Ultramafic, Klipbok complex.	02
PB 495	Ultramafic, Kouefontein.	05
1	Ultramafic near Rooiberg 2, Richtersveld (from Middlemost, 1965, p.57 , sample 1)	

(# Analyst: Dr. D.L. Reid, Department of Geochemistry, University of Cape Town.)

* Total Fe as Fe_2O_3

amphibole, chlorite, opaques white mica and carbonate) (Middlemost, 1965). The opposite situation applies in the case of the Klipbok complex.

3.2.15.2 Discussion

An understanding of the relationship between ultramafic, mafic and other rocks in the area, and their geotectonic significance is important for the following reasons:

- mafic and ultramafic rocks have an intimate relationship, suggesting a common origin and emplacement;
- they have a markedly linear pattern at the top of the Groenrivier suite. The possibility that rocks of the Klipbok complex were tectonically emplaced is therefore suggested. This general pattern is similar to that of the westerly trending ultramafics north and south of Rooiberg 2, interpreted as intruding along a line of structural weakness (Middlemost, 1965, p. 58). During a field excursion to the Klipbok complex occurrences, E.M. Moores (personal communication, June 1984) suggested that the gross layering in the serpentinites could be the product of a differentiated suite within this complex. This interpretation is justified on a mesoscale. However, taking into account that serpentinites are not confined to a particular zone within amphibolitic units and occur as spread-out entities, an interpretation favouring a separate intrusive pulse or emplacement of ultramafic magma, prior to intrusion of Vioolsdrif Granodiorite, is put forward. The timing of intrusion remains speculative as ultramafics have been segmented into "pips" during a subsequent tectonic event;
- serpentinites in the Kouefontein granite gneiss do not have clear-cut contact relationships, and no metamorphic aureole is discernable. Their location is very different to that of the Klipbok complex, viz., scattered, very few small occurrences, suggesting that they represent isolated Late Proterozoic intrusions, or simply xenoliths caught up during the emplacement of the granite. The latter interpretation is favoured if all factors including the tectonic and magmatic history are taken into account;
- the macroscale distribution of mafic and ultramafic rocks in the Richtersveld and Bushmanland Subprovinces appears to have an unsystematic pattern. However, on a smaller meso- to micro-scale the configuration of specific occurrences suggests that after initial intrusion their emplacement was tectonically controlled.

3.2.16 Correlation and intrusive relationships

The relationship between intrusive and supracrustal rocks is depicted schematically in Fig 3.19.

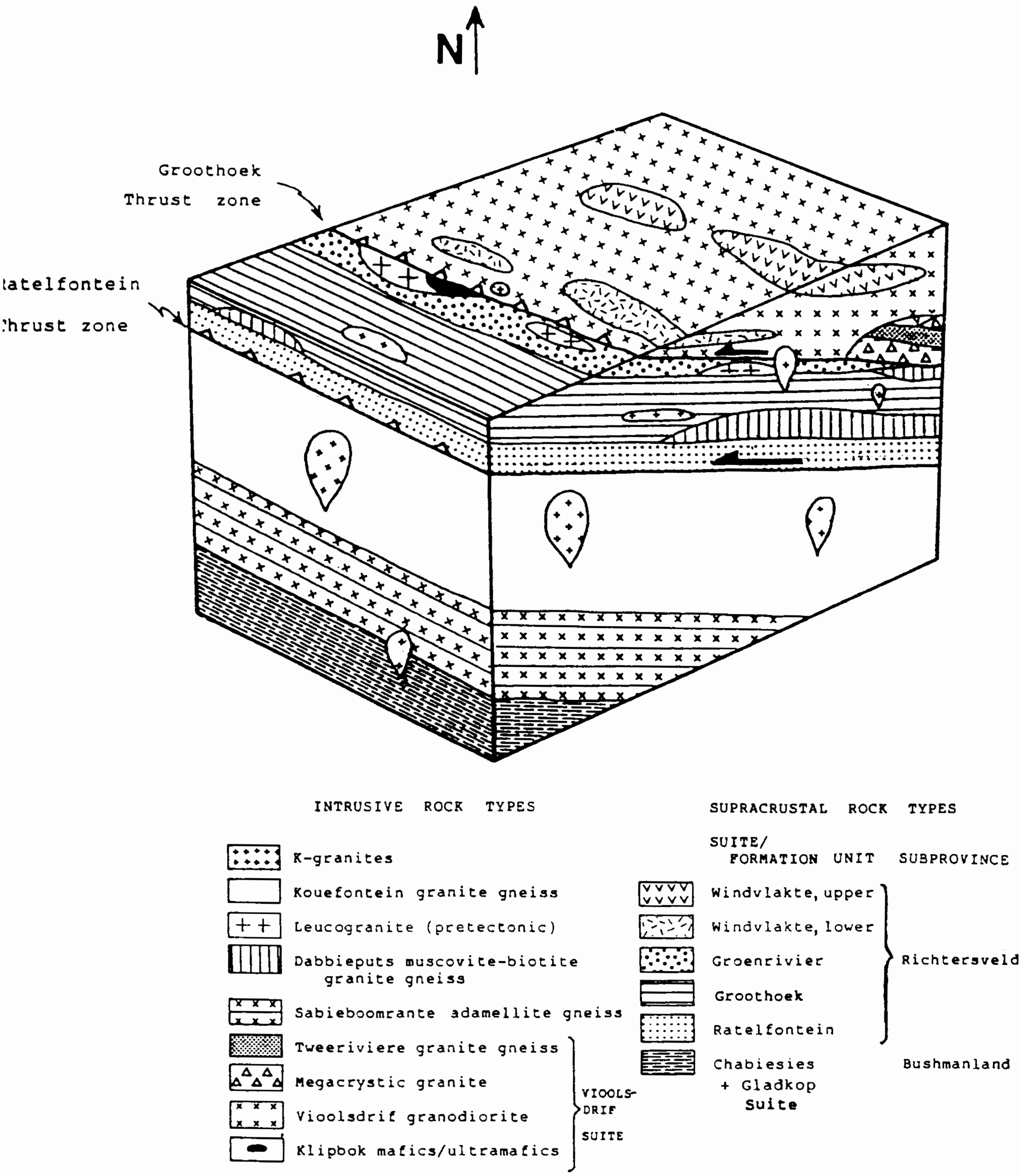


Fig. 3.19 Schematic 3-D block diagram, showing relationship between supracrustal and intrusive rocks of the Bushmanland and Richtersveld Subprovinces, in the area south-east of Eksteenfontein.

The 1900 Ma Vioolsdrif Granodiorites in the north are among the oldest intrusive rocks in the study area. They are geochemically similar to granodiorites from the type area farther north in the Richtersveld, although they appear strongly tectonised along their southern contact with supracrustals. Apophyses of this rock occur near the top contact of the Groenrivier suite.

The megacrystic granite is intimately related to Vioolsdrif Granodiorite although its occurrence, form of intrusion and geochemistry differ markedly from the granodiorites. However, field evidence suggests that the megacrystic granite is intrusive into Vioolsdrif Granodiorite. The exact relationship between this rock-type, Vioolsdrif Granodiorite and Tweeriviere granite gneiss still needs to be resolved.

The Tweeriviere granite gneiss clearly intrudes Vioolsdrif Granodiorite, and is therefore younger than the latter. The relationship of granite gneiss to the megacrystic granite is obscured through subsequent tectonism, but it is probably intrusive into the latter. Stratigraphically, the granite gneiss is placed in the Richtersveld Subprovince, but this aspect has not been fully resolved.

The Sabieboomrante and related granitoids are the subject of controversial interpretations. De Villiers and Söhnge (1959) suggest these rocks represent metamorphosed sediments, whereas Holland and Marais (1983) have interpreted rocks of similar appearance in the Okiep Terrane and near Port Nolloth as Vioolsdrif Suite- related. The current study shows that the Sabieboomrante granitoid is a thick concordant sheet of adamellitic rock intrusive into the top of the Bushmanland Subprovince, whereas other related bodies represent similar, but minor sheet-like intrusions in the Groenrivier and Groothoek suites. Geochemically the Sabieboomrante granitoid differs from typical Vioolsdrif Granodiorite in its trace element content, and therefore probably represents a different suite of rocks. Its geochemistry suggests that it may have experienced some crustal contamination. Unless otherwise proved through geochronological dating and further geochemical studies, this adamellite is interpreted as a sheet-like, S-type granitoid, intrusive into the Een Riet Subgroup. Considering the aforementioned features I propose that its age is younger than that of Vioolsdrif Granodiorite.

Leucogranite and Dabbieputs granite gneiss occur as thin sheets confined to the Groenrivier, Groothoek and Ratelfontein Suites. The leucogranite, high in Rb and low in Sr, is geochemically similar to leucogranites from farther north in the Richtersveld, but is spatially related to the mafic/ultramafic rocks of the Klipbok complex. Its relationship to the Dabbieputs granite gneiss is unknown. Correlation of the Dabbieputs granite gneiss west of Kouefontein Shear is tentative. The present interpretation is based on its stratigraphic position west of Kouefontein Shear which is similar to that east of the shear. However, further work is warranted to justify this interpretation.

The Kouefontein granite gneiss is a homogeneous thick sheet, broadest in the west and narrowing down eastwards to a point. Emplacement is visualized as taking place soon after, or during the waning stages of a thrusting event by uplifting the hanging-wall. The fact that the southern contact with the Sabieboomrante granitoid is strongly tectonised into an augen gneiss supports this interpretation. The Kouefontein granite gneiss has been emplaced at a structurally favourable position, north and south of which lithologies have divergent strike orientations. Here either an unconformity, or a major tectonic discontinuity provided the locus for magmatic intrusion. Intrusion took place prior to Groothoek thrusting as the granite gneiss has a strongly tectonised imprint resulting from this event. Small plutons of K-granite intrude the Kouefontein granite gneiss, and an abundance of pegmatites occur within it, indicating that the Kouefontein granite gneiss is older than 1000 Ma. This granite, along its eastern continuity in the Namaqua geotraverse (map code Mvl), is grouped with the Vioolsdrif Suite by Blignault et al. (1983). In this study the Kouefontein granite gneiss is interpreted as part of the Bushmanland Subprovince, and is therefore younger than the Vioolsdrif Suite. Isotopic studies are needed to verify this interpretation.

Undifferentiated discordant granite gneiss bodies occurring south of the Kouefontein granite gneiss are correlated with the Eyams granite, but this needs verification. Concordant nodular gneiss units in the Chabiesies suite are correlated with the Noenoemaasberg gneiss of the Gladkop Suite. They could also be the westernmost continuity of the Kinderlê gneiss. Banded biotite gneisses are tentatively interpreted as belonging to either the Brandewynsbank or Steinkopf gneiss of the Gladkop Suite.

K-granites, which collectively define a suite of rocks slightly older than 1000 Ma, commonly intrude the Een Riet Subgroup, the Kouefontein granite gneiss, Sabieboomrante adamellite gneiss and rarely rocks of the Vioolsdrif Suites. In the study area one small pluton intrudes Vioolsdrif Granodiorite. Their spatial distribution therefore spans both subprovinces. They differ from typical Richtersveld Suite granites in both major and trace element abundances. There is no systematic pattern to their distribution and form of intrusion, but are obviously late intrusive products. The K-granites can in all probability be correlated with the Wyepoort granite occurring east of 17°30′, because of similar field relationships. They are therefore allocated to the Spektakel Suite, but may in fact constitute a separate suite of rocks.

Amphibolites of the Klipbok complex have a restricted outcrop area, viz., at the top of the Groenrivier suite, suggesting that they were emplaced along a linear zone of structural weakness. On a mesoscale they display layering defined by grain size and mineral content. Their geochemistry suggests that they may have an island arc origin, although this proposal remains speculative. Serpentinites and hornblendites occur as entities, and are in places associated with amphibolites of the Klipbok complex. They are interpreted as separate intrusions, but could originally have formed part of a differentiated sequence within a larger mafic/ ultramafic assemblage.

The age of the Klipbok complex rocks is a problem that needs resolving. Similar rocks from farther east in Namaqualand have yielded an age of 2200 Ma (Köppel, 1980), whereas a 2000 Ma age is appropriate for some mafic rocks in the Richtersveld (Reid, 1979a). The 2000 Ma age is compatible with field relationships of the Klipbok complex. This suggests that rocks of the Klipbok complex may be intrusive into metavolcanics of the Groenrivier suite, and therefore younger than the latter. This relationship, however, is obscured by transgressive granite intrusions, e.g., the leucogranite, and subsequent deformation. The Klipbok complex is interpreted as forming part of an original island arc complex, although the present study has not been able to confirm this. They may have formed in a back-arc or fore-arc environment in which case they are probably older than the Vioolsdrif Suite.

Minor occurrences of ultramafic rocks farther south within the Kouefontein granite gneiss represent either isolated small intrusions, or xenoliths caught up during emplacement of Kouefontein granite. A favoured interpretation is discussed in Chapter 6.

3.3 LATE PROTEROZOIC INTRUSIVE ROCKS

3.3.1 Gannakouriep Dyke Suite

In the study area the dykes trend between north-northwesterly to north-northeasterly, and are located within a 30 km wide zone east of the contact with the Stinkfontein Formation. They have intruded over a considerable period of time (700 to 550 Ma), penetrating the lower-most beds of the Stinkfontein Formation, but overlain by the Nama Group (Ritter, 1980, p. 98). The dyke swarm comprises mainly mafic to ultramafic rocks, and syenitic types. They cut across pegmatites, their age being therefore younger than 1000 Ma. Currently accepted dates for the dyke swarm range between 878 to 543 Ma (De Villiers, 1968; SACS, 1980).

The dykes are considered to have originated from a mantle source and are intimately related to lithologies of the Richtersveld Suite. They were emplaced during an east-west oriented extensional tectonic event.

3.3.1.1 Mafic dykes

The majority of dykes in the study area are composed of dark green to grey mafic rocks, their grain-size varying in relation to width of intrusion. Concentrations of ferromagnesian minerals in the larger dykes give them a dark-spotted appearance.

Their width ranges between 0,5 to 30 m, averaging 5 m, whereas length varies from a few tens of metres to several kilometres. Their characteristics are mostly consistent with the hornblende-diorite dykes of Middlemost (1963, p. 171).

The dykes are coarse-grained and typically amphibolitic in character, especially those close to the contact with the Stinkfontein Formation. Blue-green hornblende porphyroblasts have a large irregular shape, enclosing small quartz grains, but do also occur as euhedral elongate grains. Subhedral to anhedral plagioclase, biotite, iron ore and epidote are accessory. In the eastern part of the study area epidote and chlorite are present in abundance.

3.3.1.2 Felsic dykes

These occur as narrow bodies throughout the area, measuring no wider than a few metres. In the field they are dark-weathering, but a fresh specimen is a grey, fine-grained rock, composed predominantly of plagioclase, iron ore and biotite. Phenocrysts of plagioclase and quartz are set in a matrix of mainly plagioclase, with minor iron ore, epidote, muscovite and biotite.

Felsic dykes here are possible correlates of the quartz syenite/quartz diorites (De Villiers and Söhnge, 1959, p. 145), and quartz bostonites (Middlemost, 1963, p. 163) from the northern Richtersveld. The rocks in the latter area are petrographically and geochemically related to the syenitic rocks of the 920 Ma Richtersveld Suite (Middlemost, 1963).

3.3.1.3 Agglomeratic dykes

One mafic dyke is different to the others in that it has a mottled appearance, due to an abundance of large feldspar and quartz porphyroclasts set in a matrix of chlorite, biotite, iron ore, epidote and carbonate. It crops out en echelon, straddling both sides of the Riethoek Shear, and forming separate north-south trending intrusions.

The plagioclase porphyroclasts measure several cm in size, and show advanced stages of replacement by sericite. Quartz clasts are well rounded, some of them show pressure shadows of recrystallized quartz and calcite. Lime-yellow epidote forms clusters within the plagioclase grains as well as small anhedral grains in the groundmass. The remainder of the groundmass comprises small plagioclase grains, biotite, chlorite and iron ore. Similar agglomeratic appearing rocks occur in the northern Richtersveld (De Villiers and Söhnge, 1959, p. 145).

Where the dyke is situated closer to the Riethoek Shear it has a strongly developed ($S(P)_2$) foliation. Here the plagioclase and quartz grains lie in the foliation and are oriented parallel to the well-developed lineation. Small polygonal recrystallized quartz grains surround the larger ones in a mortar texture. The groundmass minerals are strongly oriented in the foliation.

3.3.1.4 Field relationships

Gannakouriep dykes are readily recognized on aerial photographs as dark-toned, north-northeasterly trending linear features. They form negative-weathering topographic features such as valleys and saddles.

With regard to relative ages of the dykes, no cross-cutting relationships were noted in the field. As a generalization mafic and felsic dykes developed more or less simultaneously, based on composite relationships in a very broad Gannakouriep dyke near the Orange River (Middlemost, 1963, p. 173). Felsic dykes probably tapped a slightly shallower magma source from a contaminated product different to a deeper, more mafic magma (op. cit.). Based on this interpretation and on De Villiers and Söhnge's observations (1959, p. 138, 141) the assumption is therefore made that felsic dykes are probably slightly younger than mafic dykes.

Dykes are more prevalent in the crustal block between Riethoek and Steenbok Shears than elsewhere, becoming less abundant towards the extreme east of the study area, on Vioolsdrif Toekennings Gebied (grid reference T8, Annexure 1). East of Steenbok Shear their trend is generally north-northeast but west of this shear it is more north-south. In places there is an additional north-easterly trend. This general spread in trend is in part due to two factors:
- separate pulses of magma, each probably related to a slightly different stress field at the onset of an extended period of extensional tectonism. In the northern Richtersveld the hornblende diorite dykes have a wider distribution than the felsic dykes (Middlemost, 1963, p. 179);
- rotation of crustal blocks in response to Pan-African compressional tectonism directed from the north-northwest (see Chapter 4). This could cause the orientations of dykes and their host crustal blocks to deviate slightly from the normal trend.

In the present area shearing is evident in some of the dykes, whereas others in addition show folding similar to that described in the northern Richtersveld (De Villiers and Sohnge, 1959, p. 141).

3.3.2 Pegmatites and Quartz Veins

3.3.2.1 Pegmatites

Pegmatites here form part of a very extensive 1000 Ma pegmatite belt in the North West Cape (350 x 30 km). The pegmatite belt straddles the Richtersveld-Bushmanland high/medium grade metamorphic boundary (De Villiers and Söhnge, 1959; Nicolaysen and Burger, 1965; Hugo, 1969; Joubert, 1971; Ward, 1977). Their origin is related to fluids derived from magmatic/metamorphic processes, being emplaced mainly during the 1000 Ma tectonothermal event, with a strong structural control (Joubert, 1986a).

Mineralogically they are composed of alkali feldspar, quartz, biotite and muscovite. Orthoclase feldspar in some pegmatites differentiates them from others by imparting a readily noticeable pinkish-orange colour.

Both barren and mineralized pegmatites occur in the study area, the former homogeneous type predominating. Those containing economic minerals are virtually confined to the region north of the watershed and have been mined for beryl, feldspar and copper sulphides.

The vast majority can be recognized on aerial photographs and in the field as whitish, irregular, lens-shaped bodies. Both concordant and transecting types are present, with the former predominating. The majority of pegmatites have a north-westerly trend and a north-easterly dip. They vary enormously in size, ranging from bodies a few tens of cms wide x 1m or so long, to very large bodies about 10 m thick x 100 m long. Most of the large pegmatites occur mainly in the Groothoek suite, although there are some mineralized ones in the Vioolsdrif Granodiorite. The abundance of pegmatites in the schistose rocks of the Groothoek suite suggests that the fissile nature of these rocks probably favoured emplacement within this sequence. Tectonism has modified some of the pegmatites into boudinaged units within their host rocks.

East of 17°30′ pegmatites have intruded the "Transitional Zone" in abundance (Theart, 1980), and individual pegmatites are much larger in size than those in the present area (Ward, 1977).

3.3.2.2 Quartz Veins

Quartz veins are ubiquitous, both transgressive and concordant types being present. Many have a north-northeasterly trend approximating that of the Pan-African shear zones, especially in the eastern half of the area. They are at most 2 m wide and in several places have an associated, sporadically developed, ochrous sulphide gossan. Some quartz veins occur in zones of thrusting and have the same north-westerly trend, thus representing a late-stage precipitation in structurally favoured zones.

3.4 LATE PROTEROZOIC/EARLY PALAEOZOIC COVER SEQUENCES

3.4.1 Stinkfontein Formation

The Stinkfontein Formation forms the western boundary of the area, and no systematic study was carried out west of the contact. Detailed descriptive accounts of the rock-types are given by De Villiers and Söhnge (1959), Joubert (1971), Joubert and Kröner (1971), whereas Middlemost (1966) emphasizes their environment of deposition.

The upper Stinkfontein Formation, comprising feldspathic quartzites and sheared mafic lavas, crops out along the western boundary of the area. These rocks form part of the Vredefort Member (SACS, 1980). They have an angular relationship to the lower Stinkfontein (Lekkersing Member; SACS, 1980), and overlap the basal gneisses (Joubert, 1971). All rocks within the contact zone are highly sheared (Ritter, 1980) due to Pan-African overprinting on basement and cover rocks, making interpretation of the contact zone complex.

In a current overview of the Gariep Group in the north-western Richtersveld just south of the Orange River, M.W. Von Veh (personal communication, 1987) recognizes an upper allochthonous Marmora Terrane comprising the Schakalsberg and Oranjemund thrust sheets overlying a basal Port Nolloth Terrane. C.J.H. Hartnady (personal communication, and Excursion Guide book, 1984 Geocongress) considers that the Gariep Belt initially evolved as a rifted marginal prism on the edge of the Kalahari craton 900-700 Ma ago. A change to compressive tectonism followed. Oceanic lithosphere and accompanying passive margin assemblages were emplaced as thrust sheets over the underlying Port Nolloth Terrane composed mainly of clastic sediments and minor lavas. This is considered to have taken place until ≈600 Ma, after which Nama sedimentation proceeded within a trough created by foreland flexuring, as a result of eastward-directed Gariep tectonism. Some time after deposition of the Nama, reactivation along the former collision zone resulted in further tectonism and metamorphism.

The ≈500 Ma Kuboos-Bremen Line of acid and carbonate plutonic rocks is considered to represent hot-spot activity, and to have terminated the Pan-African tectonothermal event in the north-western Richtersveld. This took place after deposition of the Cambrian Nama Group.

3.4.2 Nama Group

3.4.2.1 General Statement

In the North West Cape the 600-500 Ma Nama Group extends from Vioolsdrif in the north to Vanrhynsdorp in the south (SACS, 1980). It unconformably overlies the Bushmanland and Richtersveld Subprovinces and occurs as isolated outliers, or sometimes as fault-bounded horsts and grabens. The Nama Group extends northwards into Namibia for some 500 km, where its northern fringes have been affected by the Damaran orogenic event.

3.4.2.2 Distribution

In the present area the Nama Group occurs in two different geologic settings, (i) unconformably overlying rocks of the Bushmanland and Richtersveld Subprovinces in the north and north-east, and (ii) a long narrow down-faulted block within the Steenbok Shear.

(i) Unconformably overlying basement rocks

The basal Kuibis Subgroup is present in the northeast of the area and forms part of the Neint Nababeep Plateau (De Villiers and Söhnge, 1959; SACS, 1980). Rock-types include whitish and grey medium-grained, recrystallized quartzites, grits and minor conglomerates, intercalated with light brown to khaki coloured shales. The quartzites sometimes have a purplish colour and in places are dark grey, reflecting an increase in proportions of iron oxide minerals. The stratigraphic thickness of this formation is at most some 80 metres (see Plate 8).

The Nama Group in this area reached a metamorphic grade transitional between diagenesis and low grade metamorphism (Ritter, 1980, p. 204).

Deformation of the Nama Group is in the form of folding and faulting associated with the Pan-African event. Folding on Vioolsdrif Toekennings Gebied in the east causes the beds to have a gentle northward plunge. Nama sediments have been steeply tilted up in zones immediately adjacent to the major north-trending shear zones similar to those farther north in the Richtersveld (De Villiers and Söhnge, 1959).

Similar deformation features occur in a small outlier of Nama rocks occurring adjacent to the Kouefontein Shear (grid reference

03, Annexure 1, and Annexure 2). Here the synformal-shaped outlier occurs about 60 m below its normal stratigraphic position.

(ii) Structurally down-faulted block

Where Nama rocks have been down-faulted within the Steenbok Shear a thin grey limestone unit is present in addition to quartzites, grits and shales. In places this limestone is silicified, and has an ochrous colour.

In general the Nama rocks of this narrow down-faulted block form a positive topographic feature. Along the contacts, however, all rocks are strongly sheared and in places highly weathered. Adjacent to the Eksteenfontein road the southernmost outcrop of Nama sediments occurs 300 metres below its normal stratigraphic position (grid reference M8, Annexure 1). Here grey, cross-bedded, recrystallized quartzites and grits dip steeply towards the east.

The Steenbok Shear truncates and forms the westernmost limit of Nama rocks in the area. Their absence west of the Steenbok Shear is the result of tectonic uplift west of this shear during the Early Palaeozoic (<500 Ma), and their subsequent removal by erosion (see Chapter 4).

A black compact nodular sulphide gossan is associated with the northern 5 km portion of the Steenbok Shear where it occurs in the form of weathered scree of fist size, and smaller fragments. Gossanous outcrops are not confined to the Steenbok Shear. Just north of the present area broad, boudin-shaped, copper-rich gossan showings are associated with major north-trending shear zones near the contact with the Stinkfontein Formation.

4 STRUCTURAL ANALYSIS

4.1 FABRIC ELEMENTS AND DOMAINS

4.1.1 Planar fabrics

Foliation, or S-surface, is used to describe any pervasive planar fabric element in metamorphic tectonites. Conventionally, where foliations of different ages clearly transect they are sequentially denoted S_1, S_2, S_3, etc. However, the study area is one showing complex overprinting relationships. For this reason the "S" notation bears a suffix, e.g. S(R), S(N) and S(P), depending on whether it formed during the Eburnian, Namaqua or Pan-African deformation events (Table 4.1). Northerly-trending $S(P)_2$, for example, is superimposed on north-westerly trending $S(N)_2$ (Fig. 4.1; Table 4.1). However, it is important to bear in mind that due to strong overprinting and virtual obliteration of Early Proterozoic structures during the 1200 - 1100 Namaqua event, interpretation of early structures is somewhat subjective. Time constraints for separate tectonic events and episodes shown in Table 4.1 are derived from isotopic evidence, as discussed in Chapter 2.

In the study area original bedding, S_0, has been obliterated by subsequent deformational events, except in one outcrop at Kliphoogte (grid reference L3; Annexure 1) where an acid tuff xenolith within Vioolsdrif granodiorite shows graded bedding. The sequence of foliation development is based on the following:

- a mylonitic fabric occurring in the Windvlakte Thrust (within the oldest supracrustal lithologies in the area) predates intrusion of Vioolsdrif granodiorite (cf. section 4.5.2.1, and see Fig. 4.32), and is therefore the oldest fabric. As this fabric is confined only to this thrust zone it is not interpreted as a regional fabric, per se, and is thus symbolized as $S(R)_1$ and not $S(N)_1$ (see Table 4.1);
- a lithological layering ($S(N)_1$) which is devoid of primary structures is folded by the earliest folds (Fig. 4.2(a)). Remnant $S(N)_1$ textures are preserved as inclusion trails in hornblende and staurolite porphyroblasts in meta-pelites of the Richtersveld Subprovince, in the east of the study area (Figs. 5.7 and 5.18). These porphyroblasts formed during the Namaqua metamorphic event at 1200 - 1100 Ma, enclosing the earlier structures (Table 4.1; and Chapter 5);
- $S(N)_2$ is the dominant regional fabric in the area. It is axial planar to F(N)2 folds which formed during the Namaqua event (Fig. 4.2(a)). $S(N)_3$ is a mylonitic fabric (in the zone of Groothoek Thrusting) which is composite with $S(N)_2$ (Table 4.1). $S(N)_3$ is best developed in the northern part of the study area. $S(N)_4$ is a spaced cleavage formed subsequent to the Groothoek thrusting episode, but composite with $S(N)_2$ (Table 4.1). It is prominent

Table 4.1 Structural elements in the study area

Deformation Event	Deformation Episode	Planar fabric element	Linear fabric element	Folds	Proposed time (Ma)
D(R) (Eburnian)	D(R)1	S_o - bedding $S(R)_1$ - foliation (in mylonite)	$L(R)_1$ (in mylonite)	not identified	2000 - 1950 (Reid, '79b)
D(N) (Namaqua)	D(N)1	$S(N)_1$ - lithological layering - Si in micro-fabric	$L(N)_1$ -intersect $S(N)_1/S_o$?	F(N)1 not identified	?
	D(N)2	$S(N)_2$ - composite and axial planar to F(N)2 folds	$L(N)_2$ - stretch-lineation	F(N)2	1200 (Barton, '83)
	D(N)3	$S(N)_3$ - sheared fabric (Groothoek Thr)		F(N)3	1100
	D(N)4	$S(N)_4$ - spaced cleavage (composite to $S(N)_2$)		F(N)4	1070 (Van der Merwe, '86)
	D(N)5	$S(N)_5$ - ext. microfabr.		F(N)5	950
D(P) (Pan-African)	D(P)1	$S(P)_1$ - foliation in P-Afr.shears	$L(P)_1$ not identified	F(P)1	750
	D(P)2	$S(P)_2$ - muscovite cleavage	$L(P)_2$ - stretch-lineation	F(P)2	680
	D(P)3	$S(P)_3$ - cleavage transects $S(P)_2$		F(P)3	530
	D(P)4	$S(P)_4$ - cleavage transects $S(P)_3$		F(P)4	480

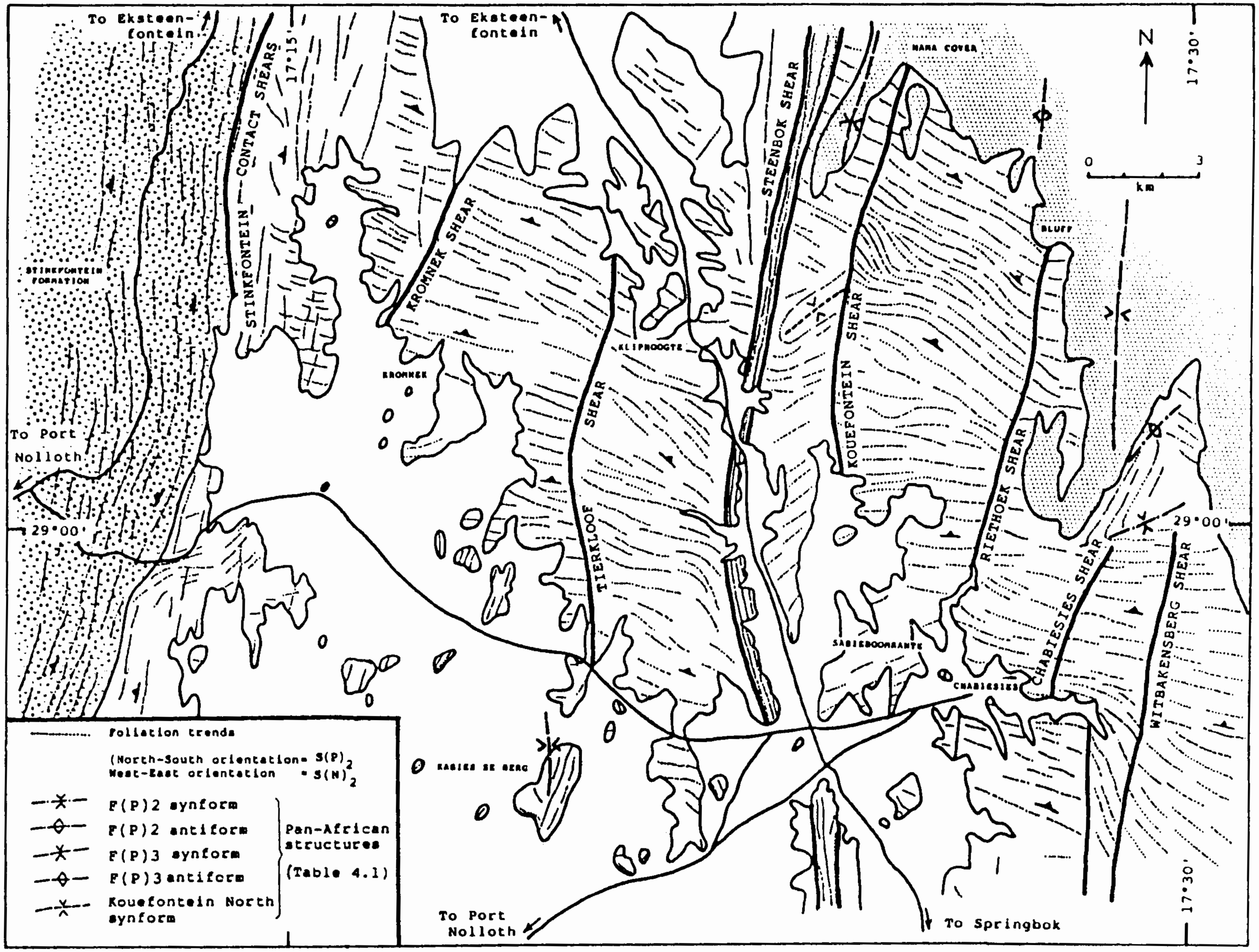


Fig. 4.1 $D(N)_2$ and $D(P)_2$ foliation trends and macrostructures in the study area.

throughout the eastern half of the study area. $S(N)_5$ represents extensional microfabric features related to northward-directed Late Proterozoic tectonism in the region (D(N)5 episode; Table 4.1);

- $S(P)_1$ is inferred to have developed in the north-northeasterly trending Pan-African shear zones which transect $S(N)_2$, almost at right angles. $S(P)_2$ is a cleavage defined by muscovite laths and quartz ribbon development in the shears (Fig. 4.19), and composite with $S(P)_1$ (Fig. 4.1). This fabric becomes progressively more pervasive in crustal blocks west of Tierkloof Shear. In the area of the Stinkfontein Contact Shears, for example, microfabric relationships show that an earlier fabric (Si) is present as inclusion trails within staurolite porphyroblasts which developed during the Pan-African event (i.e. later than the Namaqua event; see Chapter 5). Si in this case is interpreted as the $S(N)_2$ fabric (Fig. 5.2). Development of secondary foliations, "shear bands" or "cisaillement" structures across primary mylonitic foliations is particularly notable within the Pan-African shear zones. $S(P)_3$ and $S(P)_4$ are examples of such foliations which sequentially transect $S(P)_2$ (Fig. 4.20; Table 4.1).

4.1.2 Folds

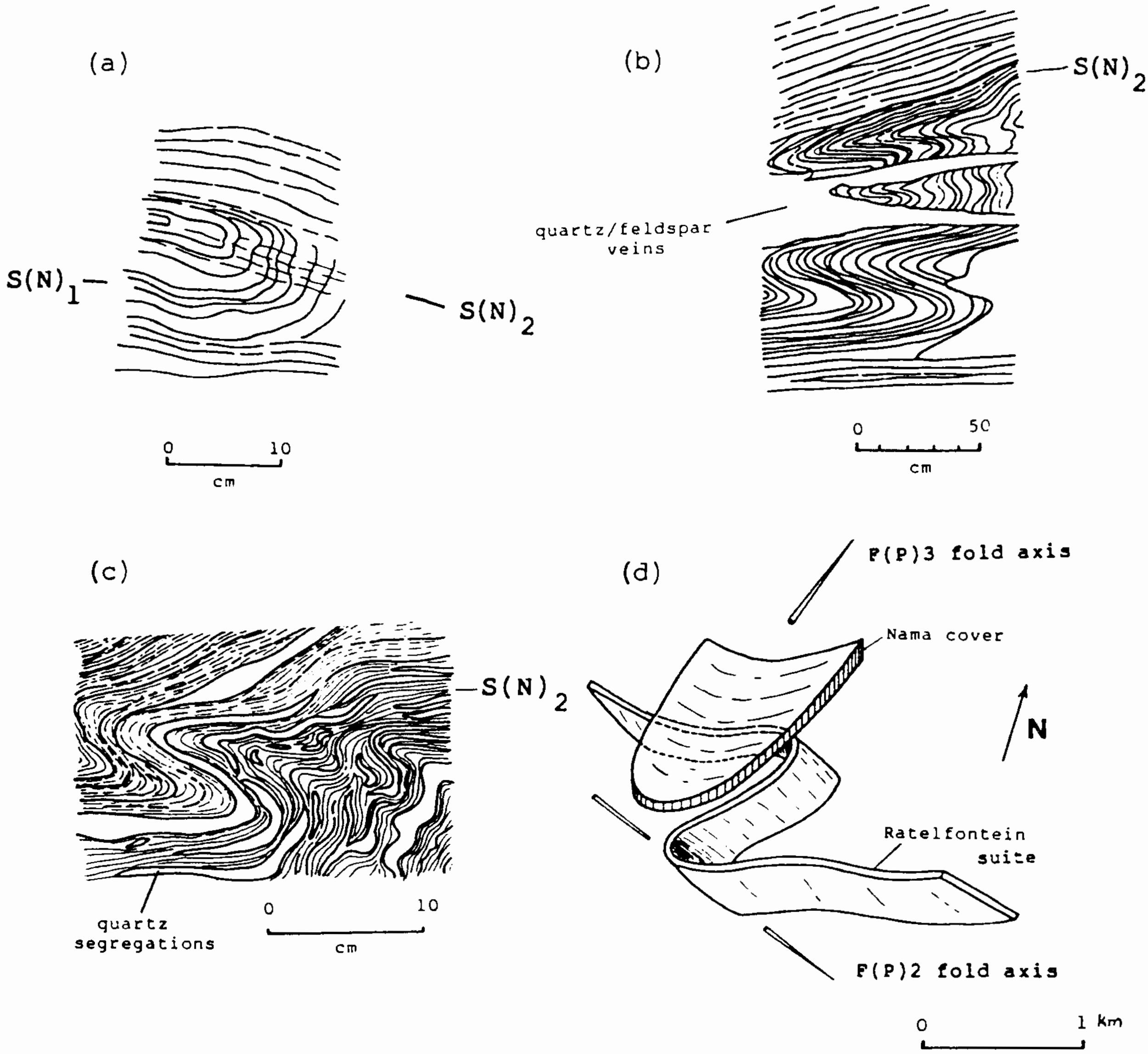
In the study area F1 folds related to an event or events prior to the Namaqua event were not identified. It can be assumed, however, that they did develop during at least one of these events. The earliest recognized folds are therefore designated F(N)2. $S(N)_2$ developed axial planar to F(N)2 (Fig. 4.2(a); and Table 4.1).

F(N)3 folds fold the $S(N)_2$ fabric, but do not develop an axial planar cleavage (Fig. 4.2(b) and (c); Table 4.1). These folds are interpreted as forming during the latter part of the Namaqua event at 1100 Ma, related to the Groothoek thrusting event (section 4.5.1.2).

F(N)4 folds are localized in minor thrusts (e.g. Chabiesies South Thrust) associated with deformation episodes which post-date the Groothoek Thrust phase (Figs. 4.51 and 4.52). These folds are interpreted as forming during the Skelmfontein thrusting episode at about 1070 Ma (Van der Merwe, 1986).

F(N)5 folds are small chevron folds in the Windvlakte Thrust which verge to the north (Plate 15). These folds formed during the last of the Namaqua- related episodes.

Similar folds of F(P)1 generation formed adjacent to the Steenbok Shear, during the Pan-African event, their fold axes aligned in $S(P)_1$ (Plate 12). The F(P)2 open fold on Vioolsdrif Toekennings Gebied (grid reference T8, Annexure 1) is interpreted as having formed in response to rotational movement on the Chabiesies Shear



(a) Isoclinal fold of F(N)2 generation (drawn from a photograph).
(b) & (c) Tight folds of F(N)3 generation (drawn from photographs).
(d) Schematic drawing of F(P)2 open and F(P)3 gentle fold on Vioolsdrif Toekennings Gebied (grid reference T8; Annexure 1).

Fig. 4.2 Fold styles in the study area.

during a subsequent deformational episode (Table 4.1; and Fig. 4.2(d)).

Tight to isoclinal folds in deformed Nama blocks within the Steenbok Shear are designated as F(P)3 folds (Plate 11). Open folds of this generation formed in the crustal blocks between the Pan-African shears. F(P)4 folds are brittle deformation structures such as kink folds which formed in the Steenbok Shear (Fig. 4.16).

4.1.3 Linear structures

4.1.3.1 Stretching lineation

This refers to the preferred dimensional orientation of inequant grains or elongate mineral aggregates formed during tectonism (Hobbs, Means and Williams, 1976, p. 271). $L(R)_1$ is defined as a stretching lineation localized in the mylonite of the Windvlakte Thrust. $L(N)_1$, defined as the intersection of $S(N)_1$ and S_0, was presumed to be co-linear with $L(N)_2$, but not actually observed in the study area. Over most of the area a stretching lineation, $L(N)_2$, is the predominant lineation (Fig. 4.3). It is represented mainly by the augen shape of quartz and feldspar grains, as well as hornblende porphyroblasts, biotite selvedges and cordierite "cigars" which are generally elongated in the down dip direction (Plate 4).

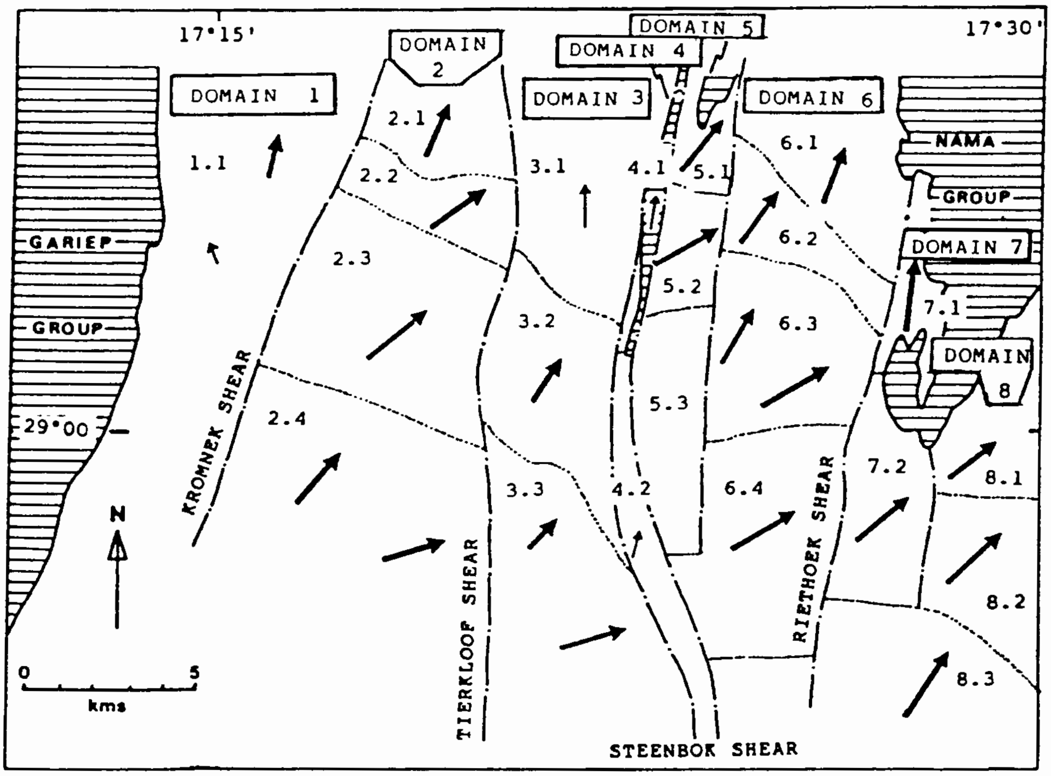
$L(P)_1$ is assumed to have formed in $S(P)_1$ during left-lateral movement along some of the Pan-African shears (section 4.6.4). Evidence for $L(P)_1$ is however lacking, and it is presumed that this lineation was obliterated during subsequent deformation and metamorphism in the shear zones (sections 4.6.4, and 5.3.3). $L(N)_2$ and $L(P)_2$ are defined according to whether they developed within the $S(N)_2$ or $S(P)_2$ foliation plane.

4.1.3.2 S-surface intersections

Relatively few outcrops showing the intersection of an axial-planar foliation and $S(N)_1$ were located (Fig. 4.2(a)). As this relationship is to be found in the hinge zone of the earlier F(N)1 isoclinal folds the resulting lineation has been designated as $L(N)_2$ (Table 4.1).

4.1.3.3 Rodding

This refers to smallish elongate cylindrical shaped bodies composed predominantly of quartz (Whitten, 1966, p. 319). It is commonly assumed that rods are parallel to local fold axes, and are genetically related to the formation of associated folds (Hobbs, Means and Williams, 1976, p. 276). Rodding is common in metapsammitic and metapelitic horizons of the Groothoek and Groenrivier



Major shear zone
Subdomain boundary
2.1 Subdomain number
Stretching lineation (pre-Pan African) $(L(N)_2)$
Stretching lineation (Pan African) $(L(P)_2)$

Arrow length is proportional to cosine of lineation plunge angle
Cos 0° =

Fig. 4.3 Structural domains and subdomains showing trend and plunge of stretching lineations in the study area.

suites, i.e. in the northernmost part of the "Transitional Zone".

4.1.3.4 Boudins

These are usually represented by elongate or chocolate tabloid structures of competent material within a less competent sequence. They are frequently developed in the Ratelfontein, Groothoek and Groenrivier suites. Boudins are regarded as structures resulting from local extension and can readily be reproduced in the laboratory (Hobbs, Means and Williams, 1976, p. 283).

4.1.3.5 Slickensides

Slickenside lineations are uncommon, but are present in some chloritic schist units. In general, they were found to be unreliable for deciphering the sense of movement direction.

4.1.3.6 Fold axes

Orientations of fold axes of the following styles of folds were measured:
- isoclinal and tight folds. These mostly represent the earliest phase of folding and have their fold axes plunging pre-dominantly towards the north-east;
- open and gentle folds. Fold axes, as a generalization, plunge towards the north and modify the geometry of earlier folds;
- microfolds. Crenulations of microfolds, which are more common in pelitic rock-types and phyllonites, were measured as lineations.

The most useful criteria for correlating tectonic structures and events across this area are transecting foliation relationships, fold style and orientation. Park (1969) and Williams (1985) have critically assessed the validity of this approach and warn of complications arising if these criteria are applied to correlation between structurally complex areas, especially where outcrops are not continuous.

4.1.4 Structural Domains

For the purposes of a detailed structural analysis the study area was divided into eight structural areas (domains), the boundaries of which are defined by major north-south trending shear zones (Annexure 3). Subdomains were outlined by homogeneous geometric patterns (defined by foliation and lineation) within each domain.

Judicious selection of subdomains by trial and error can enhance interpretation of planar and linear fabric orientations. The three-dimensional geometry of these structural elements is more easily envisaged, and kinematic interpretations can be applied

(Turner and Weiss, 1963, p. 8).

The following generalizations apply with regard to S-surfaces and lineation orientations:

- $S(N)_2$ in the majority of subdomains strikes west-north-westerly and dips north-northeasterly. Macro-folding of $S(N)_2$ is relatively rare although there is a large open fold on Vioolsdrif Toekennings Gebied (Subdomain 8.1, Domain 8, Annexure 3; and grid reference S8, Annexure 1). $S(N)_2$ in the area of the Groothoek Thrust has a consistent strike (290°) and dips about 45° north-northeast (Subdomains 1.1(a), 2.1, 6.1 and 7.1, Annexure 3). Lineations here consistently plunge down-dip (023°/43°), as do the fold axes of small tight folds (Subdomain 6.1, Annexure 3). This pattern is related to the Groothoek Thrust zone which separates competent lithologies of the Vioolsdrif Suite from less competent supra-crustals;
- south of the Groothoek Thrust zone $S(N)_2$ orientations vary and differ slightly from those to the north. $S(N)_2$ generally dips more steeply in the north than in the southern part of the area. $L(N)_2$ stretching lineations plunge towards the north-east in the south, and to the north-northeast in the north (Fig. 4.3). Tight F(N)3 axes vary in plunge directions from north-west to easterly (see Subdomain 6.2, Domain 6, Annexure 3);
- $S(P)_2$ and $L(P)_2$ are mainly localized in and around the broad, major shear zones, e.g., Stinkfontein Contact Shears (Sub-domain 1.1b, Domain 1; Annexure 3) and Steenbok Shear (Subdomains 3.1, 4.1 and 4.2; Annexure 3);
- left-lateral movement along some of the major shear zones has in places caused a change in the orientation of $S(N)_2$, e.g. Subdomain 5.3, Domain 5, Annexure 3).

4.2 SHEAR ZONES AND CRITERIA FOR MOVEMENT DIRECTIONS

4.2.1 Deformation in shear zones

Shear zones are local regions of material softening in the earth's crust, usually developed as discrete quasi-planar structures, but often as anastomosing systems throughout a deformed zone (Bell, 1978; Bell and Hammond, 1984). The actual scale of deformation can vary from submicroscopic shear bands and slip planes to broad tectonic zones some tens of kilometres wide (White et al.,1980). Shear zones are common, especially in Precambrian terranes, with Namaqualand and the Richtersveld being no exception (De Villiers and Söhnge, 1959; Joubert, 1971; Ritter, 1980; Blignault et al, 1983).

Sibson (1977) proposes a comprehensive conceptual model which essentially recognizes that shear zone rocks respond differently in the plastic lower crustal zone, compared to the brittle upper crustal zone (Fig. 4.4). The transition zone between these two

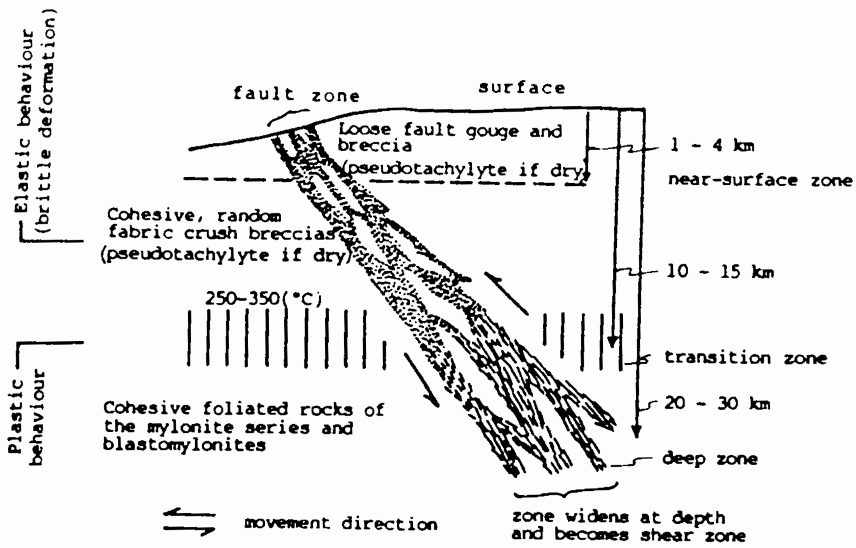


Fig. 4.4 Conceptual model of a shear zone (after Sibson, 1977, Fig. 8).

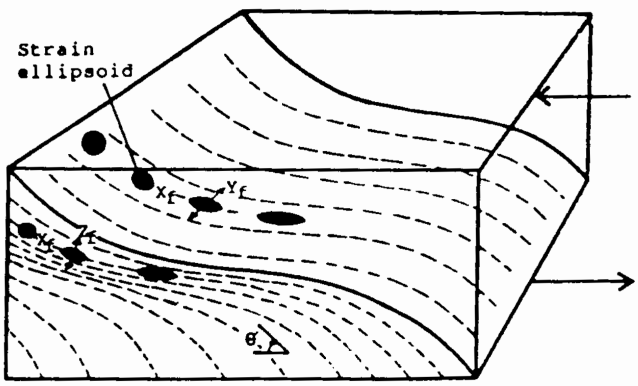


Fig. 4.5 Fabric in a ductile simple shear zone (after Ramsay, 1980, Fig. 9).
θ' is the angle between the long axis of the strain ellipsoid and the shear direction.
X_f, Y_f and Z_f represent maximum, intermediate and minimum strain directions.

levels lies between 10 to 15 kilometres, approximating greenschist facies metamorphic conditions. Above the transition zone rocks within a shear zone are chiefly breccias with random fabric, a fault gouge being encountered near the surface. Below the transition zone foliated mylonites and blastomylonites develop through mechanisms involving intense recrystallization, usually in a ductile shear widening in depth. The main criteria for distinguishing sheared rocks originating at deeper crustal levels from those developed in shallower zones, is their degree of foliation, and cohesion of the matrix material.

Within zones of simple and pure shear deformation, planar and linear fabrics are systematically related to the finite strain ellipsoid (Ramsay, 1980). When the Y axis remains parallel to the walls of the shear zone and unaltered in length during shearing, the condition is known as plane strain. The X-axis is rotated towards the direction of tectonic transport and elongated, whereas the Z-axis is shortened (Fig. 4.5). The planar and linear fabric orientations therefore track the orientation of the principal finite strain axes, but axial ratios of the strain ellipsoid can be obtained from the angle between the schistosity and the shear zone boundaries only if there was no initial fabric in the rock prior to shearing (Ramsay and Graham, 1970).

Pre-existing linear structures are rotated towards the X axis of the strain ellipsoid (Bell, 1978). Initially cylindrical fold axes can be rotated so that they plunge in the down-dip direction, or become curved to form sheath folds (Minnigh, 1979; Ramsay, 1980).

4.2.2 Development of mylonites and cataclasites

The original definition of mylonite (Lapworth, 1885) is simply that of a banded microbreccia with fluxion structure. Subsequent definitions (Bell and Etheridge, 1973; Tullis et al., 1982) take into account the crustal level of deformation. There is general agreement that, if rocks are deformed at deeper crustal levels, then grain refinement by syntectonic recrystallization and neomineralization of the rock produces mylonites (White et al.,1980). If deformation takes place at higher crustal levels, then cataclasites form by brittle fracturing (Higgins, 1971).

There is still controversy about the mechanisms that produce mylonites. Some regard mylonite development as a simple shear phenomenon (Ramsay and Graham, 1970; Bell, 1978). Others envisage a flattening process by pure shear which produces an elongation parallel to the mylonite banding (Johnson, 1967; Sinha Roy, 1977).

The "protomylonite-mylonite-ultramylonite" series refers to a progressive decrease in size and number of porphyroclasts within a sheared matrix. This series represents sequential deformation products resulting from increased deformation (Higgins, 1971; Christie, 1960). Mylonites, for example, contain between 50-90% matrix material. A recrystallization of the original mylonite texture results in a "blastomylonite" - where it is difficult to recognize the original mylonitic texture (Higgins, 1971).

Mylonite microfabric development is inferred to involve grain-size reduction, continuous recrystallization, ribbons (usually composed of quartz) kink bands, microfaulting and cracking (especially in feldspars), and the development of myrmekite. The responsible mechanisms include dislocation gliding, dislocation creep, diffusion creep (pressure solution), grain boundary sliding, and recovery usually seen in subgrain recrystallization (Tullis et al., 1982). Deformation may occur under changing physical conditions within a particular zone of movement, e.g., shearing through different crustal levels may progressively move a particular material element of mylonite through different temperature, stress, and effective pressure gradients, resulting in microstructures that are difficult to interpret (Tullis et al., 1982).

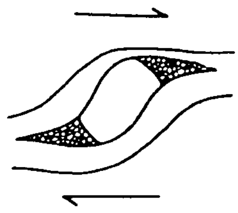
Different mineral responses to increasing grades of metamorphism is reasonably well known (Theodore, 1970; Tullis et al., 1982; Simpson, 1985). Quartz deforms plastically at lower greenschist facies whereas K-feldspar and plagioclase show only brittle deformation. The latter minerals require lower amphibolite facies conditions to deform plastically. Phyllosilicates kink rather readily at the lower grades of metamorphism.

Fluid plays a significant role in shear zones by effectively weakening the rock, enhancing pressure solution, and providing more fluid phases for metamorphic reactions during deformation (Tullis et al., 1982). During dilatancy within a shear zone large volumes of fluid can pass through and give rise to quartz veins and pegmatites.

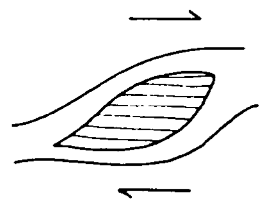
4.2.3 Sense of movement directions

Simpson and Schmid (1983), Simpson (1984), Lister and Snoke (1984) and Takagi (1986) successfully employ a number of petrographic criteria to deduce movement directions in sheared rocks (Fig. 4.6). The most useful are:

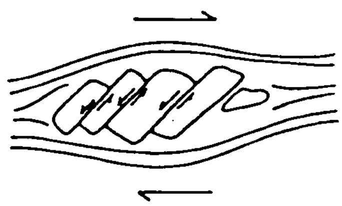
(i) Asymmetric augen structures, usually shown by feldspar, quartz and mica. Grains such as quartz are often recrystallized in pressure shadow areas, enhancing the asymmetric shape of the larger grain. Micas deformed under low grade metamorphic conditions may be modified into a fish shape through boudinaging and microfaulting of pre-existing grains (Fig. 4.6(a) and (b)). If, however, the porphyroclast has experienced a complex deformation history then reliable interpretations of sense of movement are not always



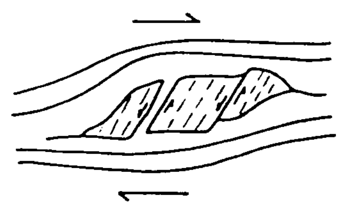
(a) Asymmetric shape to quartz grain. Recrystallized "tails" are composed of same material as host porphyroclast.



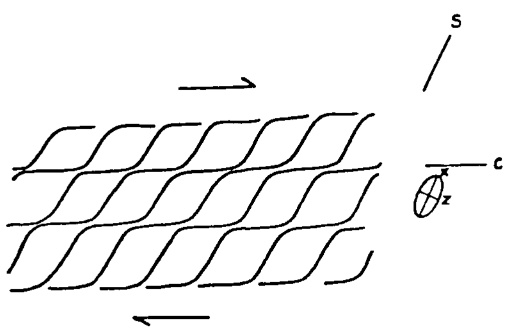
(b) Asymmetric fish-shaped mica grain.



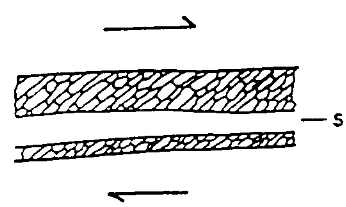
(c) Displaced broken grains of feldspar, for example.



(d) Fractured mica grains in ductile matrix.



(e) Development of S and C surfaces.



(f) Grains recrystallized with their long axes oblique to the foliation, S.

Fig. 4.6 Criteria used to determine sense of movement in sheared rocks. (a) to (d) & (f) after Simpson and Schmid (1983). (e) after Lister and Snoke (1984).

possible (Passchier and Simpson, 1986);

(ii) Displaced broken grains, where the movement sense along microfractures in brittle feldspar and mica, for example, opposes that producing the asymmetric grain shape (Fig. 4.6(c) and (d));

(iii) S (schistosity) and C ("cisaillement" shear band) surfaces. S-surfaces are defined by the preferred orientation of biotite, quartz and feldspar aggregates. C-surfaces are discrete slip surfaces of high strain, always less steeply dipping than S, usually defined by the preferred alignment of mica grains (Fig. 4.6(e)). The younger foliation, C, rotates into parallelism with S and becomes closer spaced where shearing has been most intense (Berthé et al., 1979; Bell and Hammond, 1984);

(iv) Grains recrystallized with long axes oblique to the foliation (Fig. 4.6(f)), produced by continuous dynamic recrystallization within mylonites;

(v) Crystallographic fabric asymmetry. Preferred orientation of quartz C-axes, e.g., can be used to determine movement sense and direction within a particular tectonite (Turner and Weiss, 1963, p. 230). This technique was not employed in the present study because ambiguous interpretations of the results are possible (Eisbacher, 1970; Simpson and Schmid, 1983).

4.3 PAN-AFRICAN SHEAR ZONES AND RELATED STRUCTURES

4.3.1 General statement on the orientation of shear zones, and calculation of displacement

The major Pan-African shear zones strike slightly east of north and normally dip near vertically or steeply towards the west. Those described here are at least 30 m wide, spaced on average 4 km apart. Relative displacements are recorded by the offset of the Ratelfontein suite, the most continuous stratigraphic marker unit in this area (Fig. 4.7). Generally, displacements on the shears increase from east to west.

For each shear zone, displacements are calculated using the following method (Ragan, 1968, p. 130):

(i) Each shear zone is regarded simply as a fault offsetting the north-westerly striking, north-easterly dipping strata. A plan drawing is made showing the average orientation of the fault and offset strata either side of the fault;

(ii) The orientation of each shear zone and of two differently oriented sets of strata on either side are plotted on a stereogram, so that pitch angles in the fault plane can be measured. In

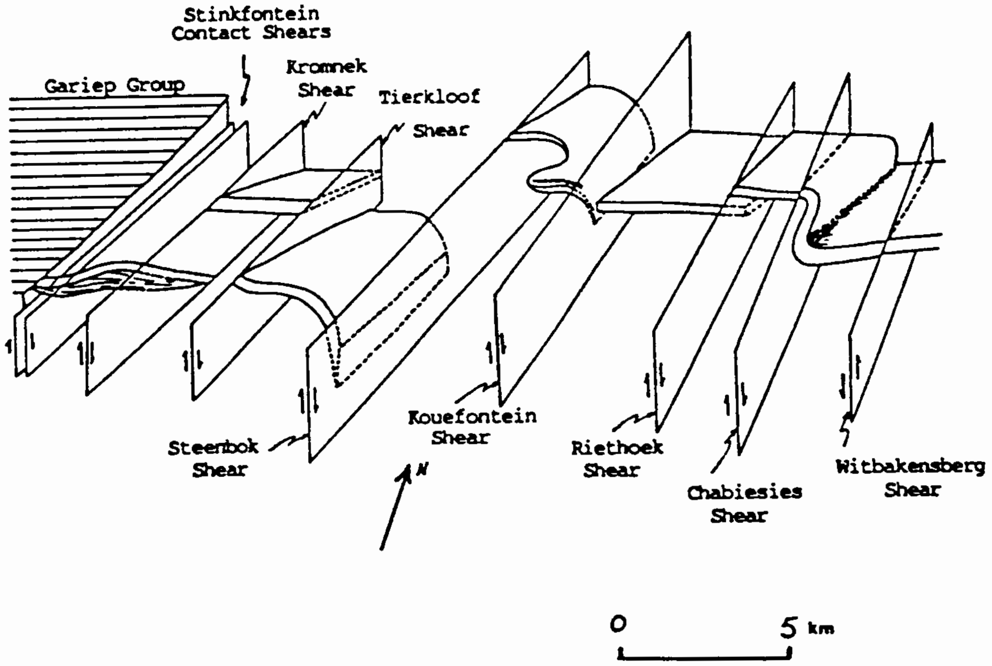


Fig. 4.7

Schematic diagram showing orientation of Pan African shear zones, and offset positions of the Ratelfontein suite in the study area. Arrows indicate sense of movement during the main high-angle thrusting episode which took place after a proposed strike-slip movement along the Kromnek and Steenbok shears.

most cases the base of the Ratelfontein suite and top contact of the Sabieboomrante adamellite gneiss abutting the fault are used as reference horizons. The trend and plunge of lineation in the shear is also plotted;

(iii) On a fault-plane section, pitch angles are used to plot the traces of two strata which intersect at two "piercing points" on either side of the fault. The lineation pitch angle is also plotted from one or both of these points. X refers to the piercing point on the west section of the fault whereas Y corresponds to that on the east section of the fault;

(iv) Net-slip (X-Y) is measured in the fault-plane as the vector connecting the piercing points, which were contiguous before faulting/shearing occurred;

(v) The angle of rotation is measured on the stereogram, after rotating the pole of the fault to the centre of the stereogram, and measuring the angular distance between the poles of strata after they have been moved by the same amount as the fault, and in the same direction;

(vi) The centre of rotation is plotted in the fault plane by subtending the angle of rotation as the apex of an isosceles triangle on a base X-Y (net-slip), in cases where this is applicable.

4.3.2 The Major Shear Zones, and Associated Structures

4.3.2.1 Witbakensberg Shear

This is located in the extreme east of the area (Annexure 1 and 3). It strikes 190°, dips 88°W, and shows a lineation oriented 340°/80°. At its northern end the shear zone passes into a fault breccia cemented with limonite. A lesser amount of displacement here suggests that there is a rotational component along the shear zone with variable, southwards increasing displacement. The fault hinge-line is probably situated just to the north of the Ratelfontein suite outcrop (grid reference U8, Annexure 1).

Marker units, namely, two closely spaced quartzite horizons intercalated in gneisses of the Chabiesies suite, show that the strata are down-thrown to the west by some 300 metres (Fig. 4.8). A plot of piercing points on a fault plane section suggests that an angle of rotation of about 5° satisfies the displacement of strata on the west side of the fault (Fig. 4.9).

The Sabieboomrante adamellite gneiss has a broader outcrop just west of this shear, suggesting that the granitoid is thicker in the up-dip section.

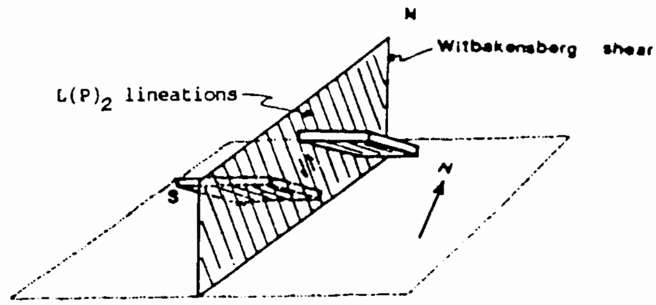


Fig. 4.8 Schematic diagram showing displacement of marker bed along the Witbakensberg Shear. Arrows in the shear zone show sense of movement along the $L(P)_2$ lineation (strata on the west side of shear have been dropped down relative to those on the east).

4.3.2.2 Chabiesies Shear

This shear is only some 5 km long, strikes 025° and dips vertically. Although schistose and amphibolitic xenoliths of the Chabiesies suite abut the shear on both sides, displacement near its southern extremity is evidently minimal. However, the large synformal structure outlined by the Ratelfontein suite on Vioolsdrif Toekennings Gebied along its northern extremity may owe its origin to this shear, in which case displacement along the fault may be increasing towards the north. The hinge point of the shear is then just north of the Chabiesies homestead.

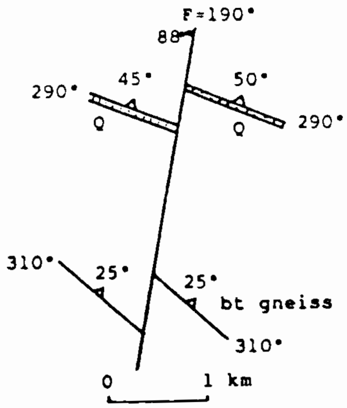
4.3.2.3 Riethoek Shear

The shear zone has a strike of 190° and a 65° westerly dip along its southern portion (see Plate 9), but steepens up northwards (Fig. 4.10). $L(P)_2$ has the orientation 297°/64°.

Schistose and amphibolitic xenoliths of the Chabiesies suite abut its eastern side only (grid reference Q8 and 9, Annexure 1). Where the shear zone transects the Ratelfontein suite quartzite beds have been rotated to plunge steeply within the shear itself.

The sheared Kouefontein granite is seen to be composed essentially of quartz and muscovite (Fig. 4.11). Quartz occurs as ribbons while needle-like laths of muscovite align parallel to the $S(P)_2$ foliation.

The western block has moved up relative to the eastern block on a high-angle thrust. Microscopic evidence in the form of asymmetric



(a) Plan of Witbakensberg Shear showing displacement of quartzite (Q) and biotite gneiss (bt gneiss) units.

(b) Stereogram data.

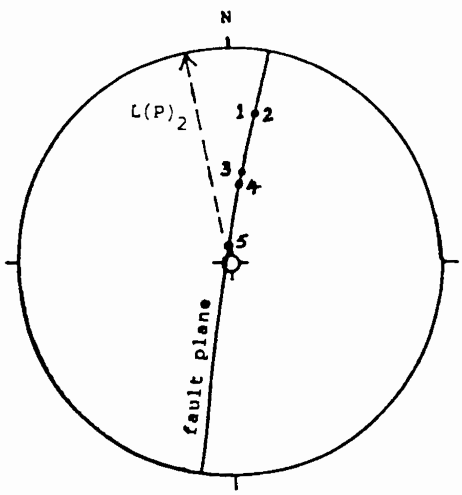
1 = pitch of biotite gneiss west of fault (22°)

2 = pitch of biotite gneiss east of fault (22°)

3 = pitch of quartzite west of fault (44°)

4 = pitch of quartzite east of fault (48°)

5 = pitch of $L(P)_2$ in fault (80°)



(c) Displacement in fault plane.

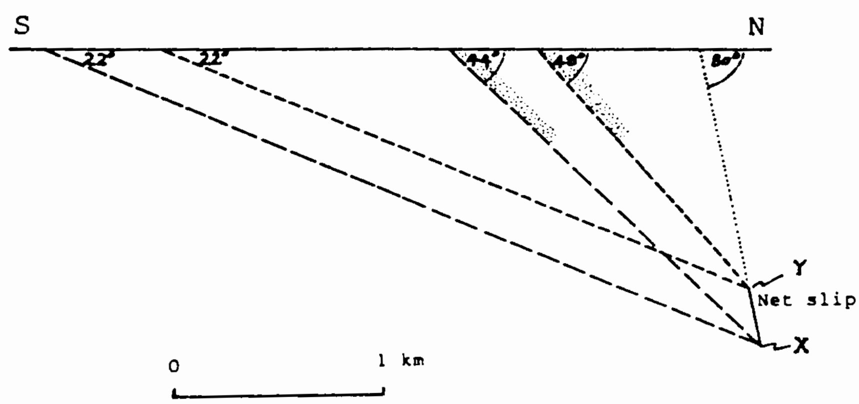


Fig. 4.9 Data and methods used to calculate displacements on Witbakensberg Shear.

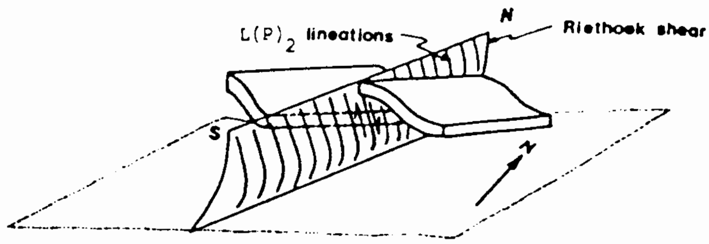


Fig. 4.10

Schematic diagram showing left-lateral offset position of Ratelfontein suite strata along the Riethoek Shear. Arrows in the shear zone show sense of movement parallel to $L(P)_2$ i.e. strata on the west side of shear have moved upwards during the last tectonic episode.

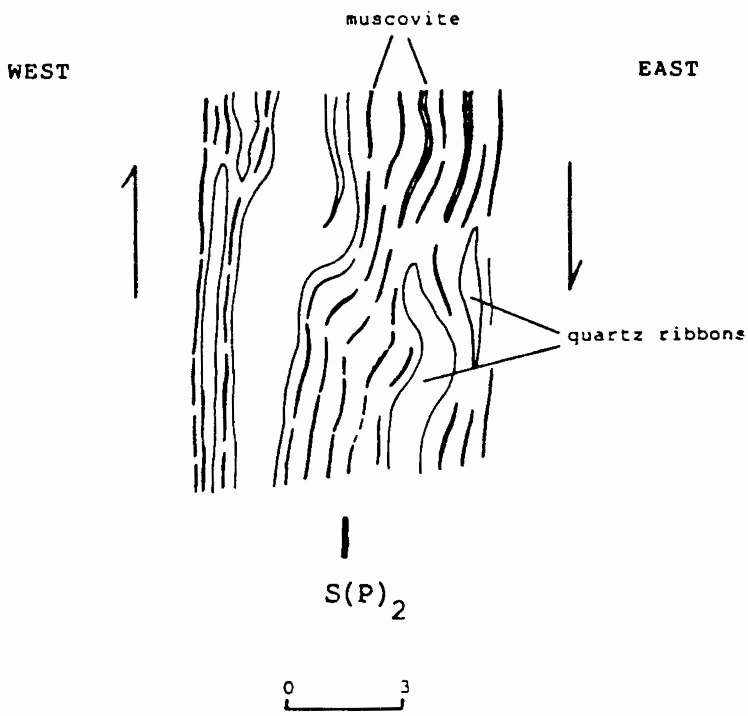


Fig. 4.11

Microstructures in muscovite quartz schist, Riethoek Shear. Muscovite outlines $S(P)_2$ foliation. Asymmetric shape of quartz ribbons shows that west block has moved up. Drawn from a photomicrograph.

shape to quartz ribbons (Fig. 4.11) confirms this sense of movement. Pressure shadows developed around quartz inclusions in an agglomeratic Gannakouriep dyke adjacent to this shear zone, give this same movement sense.

A net-slip of 7.5 km is measured on the fault plane section, with a 20° angle of rotation satisfying the geometrical configuration of strata either side of it (Fig. 4.12). This fault solution indicates that the crustal block to the west of the Riethoek Shear has been uplifted and rotated so that the strata on that side have a shallower dip than those to the east. The Riethoek Shear is therefore a high-angle thrust fault with a rotational sense of movement.

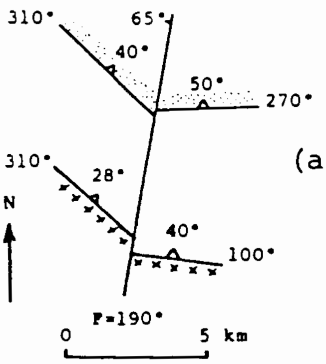
There are two important assumptions in this technique. The first is that the upper contact of the Sabieboomrante adamellite gneiss and lower contact of the Ratelfontein suite, when projected down dip, always intersect at a point. This means that the Kouefontein granite gneiss, located between these two contacts, pinches out in depth. Although this may be true, it cannot be verified at present. The second assumption is that the pronounced stretching lineation ($L(P)_2$) represents the slip direction in the shear zones. This seems to hold in the case of the Riethoek Shear, but in some other cases, e.g., the Steenbok Shear, it may not be true.

4.3.2.4 Kouefontein Shear

This has an average strike of 010°, dips vertically, and shows an en echelon pattern, especially at its southern end. Where it crosses from the Kouefontein granite gneiss to the Sabieboomrante adamellite gneiss it swings to the south-east (grid reference N7 south eastwards to P9, Annexure 1).

The western crustal block has rotated by 3° relative to the eastern block, about a centre of rotation some 3 km south of the northern area margin (Fig. 4.13). Within the western crustal block the basal unconformity of the Nama Group has been downdropped by at least 100 m in the north, while the Ratelfontein suite has been uplifted by 60 m some 4 km farther to the south.

A remarkable display of the modification of earlier structures south of the centre of rotation can be seen in the deformation of the Kouefontein synform (grid reference N4, Annexure 1). This structure is a north-easterly plunging synform in the quartzites of the Ratelfontein suite. Approaching from the west, its plunge steepens towards the shear zone, until the quartzites assume a vertical position within the shear zone itself (see Fig. 4.14).



(a) Plan of Riethoek Shear showing displacement of Ratelfontein suite (dots) and Sabieboomrante adamellite (crosses).

(b) Stereogram data.

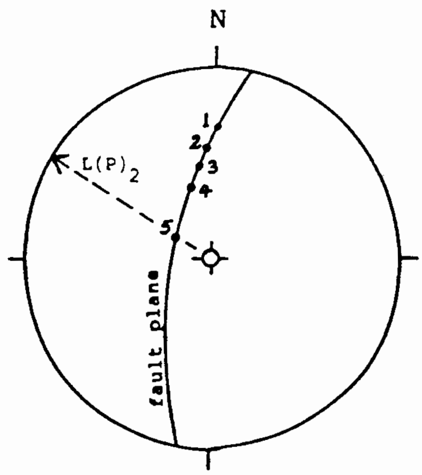
1 = pitch of adamellite contact west of fault (24°)

2 = pitch of Ratelfontein suite contact west of fault (34°)

3 = pitch of adamellite contact east of fault (43°)

4 = pitch of Ratelfontein suite contact east of fault (55°)

5 = pitch of $L(P)_2$ in fault (84°)



(c) Rotational displacement in fault plane.

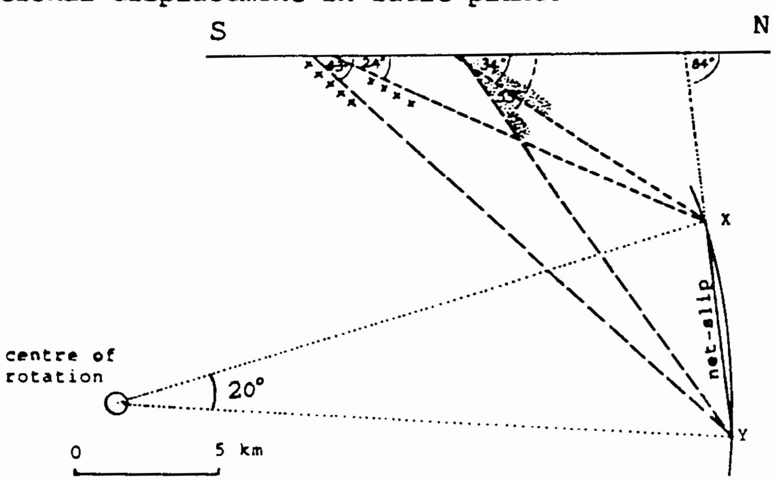


Fig. 4.12 Data and methods used to calculate displacements on Riethoek Shear.

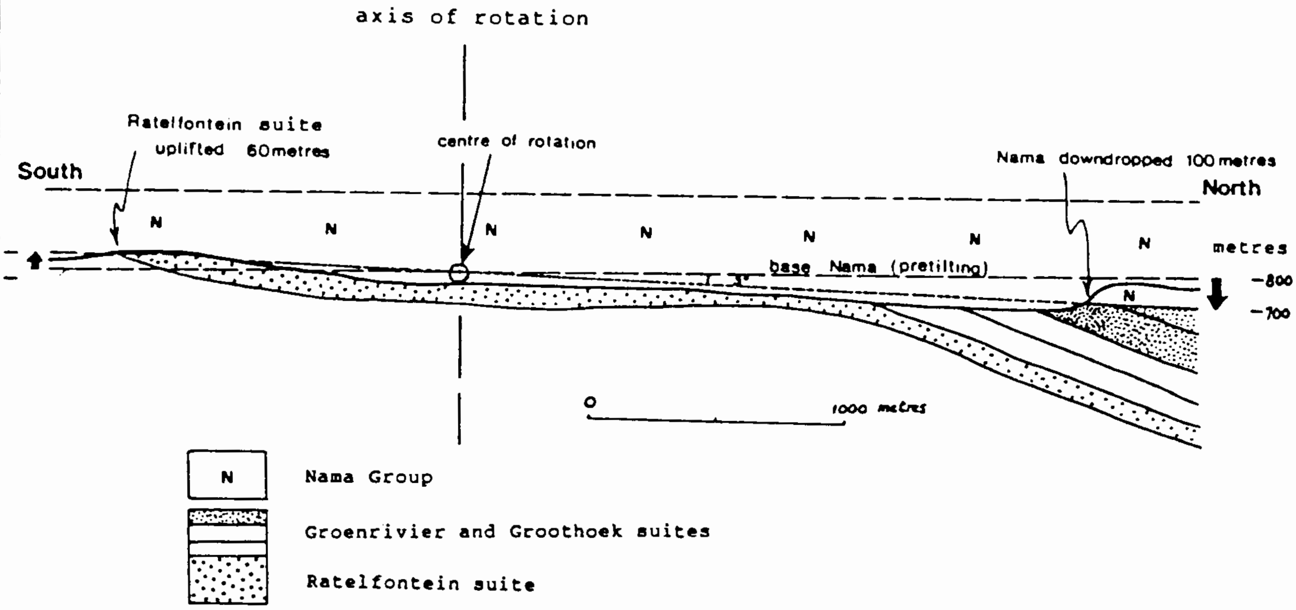


Fig. 4.13 North-south section along northern portion of Kouefontein Shear, showing rotational movement of the western crustal block.

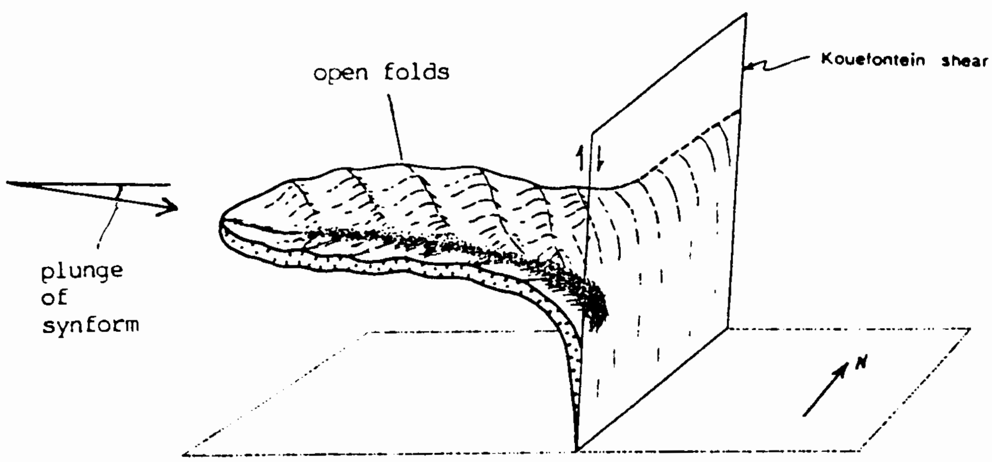


Fig. 4.14 Schematic diagram of the Kouefontein Synform. Note steepening of strata eastwards until a vertical orientation is reached in the Kouefontein Shear zone.

4.3.2.5 Steenbok Shear

4.3.2.5.1 Macro features/structures

This is the most conspicuous shear zone in the area. A generally positive weathering feature, it is much broader than the other shear zones, averaging some 200 m wide. It is prominent along almost its entire outcrop length, and is clearly visible on aerial photographs (Plate 10).

Joubert (1971) originally described this shear zone from the area farther south, noting particularly its left-lateral displacement, modification of quartzitic units, the characteristic development of phyllonites and associated kink folding. A major fracture in the coastal area of Namaqualand, it is traceable for at least 65 km from the farm Drooge Kraal, south-east of Port Nolloth, into the Richtersveld.

From south to north in the present area, its strike changes from slightly west of north to slightly east of north. The average orientation of $S(P)_2$ foliations and $L(P)_2$ lineations within this shear, and its changing trend, are more clearly depicted in structural Domain 4 (see Annexure 3).

Along and within the northern part of the Steenbok Shear, a sliver of Nama Group rocks has been displaced by at most some 300 m below their normal stratigraphic level. Nama strata within this down-faulted slice are tightly folded and steeply dipping (Fig. 4.15b).

4.3.2.5.2 Mesostructures

(i) S(P) foliations

Where the shear zone traverses quartzitic rocks these have behaved plastically by folding, maintaining their continuity. Otherwise coarse-grained quartzo-feldspathic rocks, such as those of the Kouefontein granite gneiss and Sabieboomrante adamellite gneiss, have been transformed into phyllonites with a very well developed $S(P)_3$ and $S(P)_4$ foliation. These consist essentially of quartz, muscovite, chlorite and iron ore, with variable amounts of sericite and occasional calcite. The latter retrograde minerals are preferentially developed within certain narrower zones. Iron ore shows two generations, the first strung out along the foliation, the second in the form of euhedral crystals clearly younger than the chlorite and sericite. The Steenbok Shear phyllonites incorporate thin slices of white mica-bearing quartz schist which may be remnant slices of a different protolith, or may be products of chlorite-consuming reactions at low temperature/high pressure metamorphic grade.

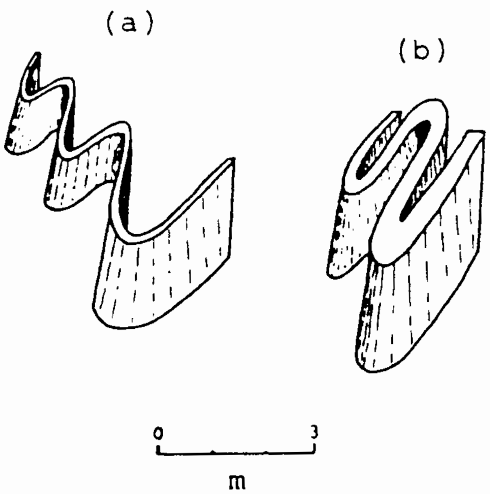


Fig. 4.15

Fold styles exhibited by competent beds within the northern portion of the Steenbok Shear (F(P)3 generation.) (a) open, (b) isoclinal.

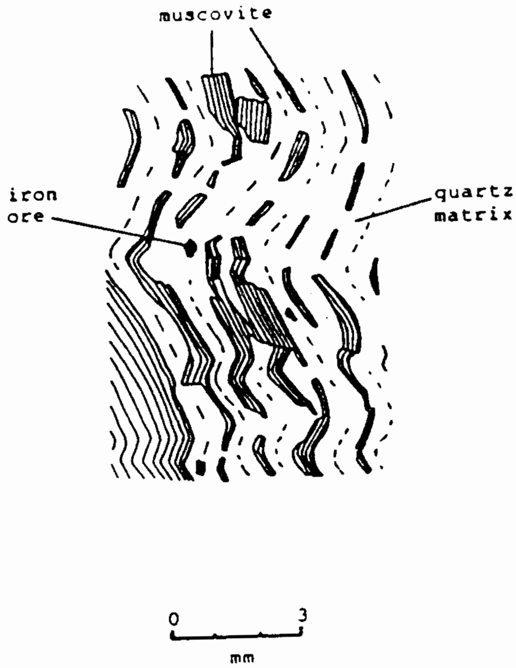


Fig. 4.16

Kink folds developed in phyllonite, Steenbok Shear (F(P)4 generation.) Traced from a photomicrograph.

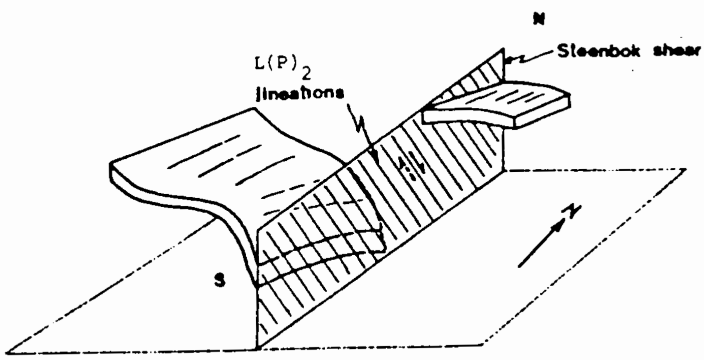


Fig. 4.17 Schematic diagram showing left-lateral offset position of strata along the Steenbok Shear. Arrows in the shear zone indicate a later movement direction parallel to $L(P)_2$ during which strata on the west side moved upwards.

(ii) Folds

The most prominent mesofolds (F(P)3; Table 4.1) occur within the northern part of the downdropped Nama Group slice. These structures vary from an open style nearer the western contact, to isoclinal in the central part of the shear zone (see Fig. 4.15, and Plate 11). In all cases the fold axes plunge steeply south-southwest, indicating that a left-lateral sense of movement produced them. $S(P)_3$ is always developed as an axial planar foliation to these folds.

An outcrop along the western contact of the Steenbok shear shows south-west plunging, s-asymmetric, tight folding of a deformed Gannakouriep dyke (grid reference M8, Annexure 1), developed during left-lateral movement. Deformation of Gannakouriep dykes in the Richtersveld has been mentioned previously by De Villiers and Söhnge (1959), Joubert (1971) and Ritter (1980). The age of this deformational episode can therefore be placed as at least younger than 878 Ma.

Immediately east of the shear zone, folds of similar (Ramsay Class 2) style (F(P)1 generation, Table 4.1; and Plate 12), showing slip in a direction parallel to their axial planes, were located within the Sabieboomrante adamellite gneiss (grid reference M8, Annexure 1). These similar folds developed under ductile deformation conditions during the initial stages of the Pan-African tectonic event, their fold axes aligned parallel to the Steenbok shear zone.

Kink folds (F(P)4) represent brittle structures mainly within the Steenbok Shear (Fig. 4.16). Their fold axes are horizontal.

(iii) Lineations

The most prominent lineation in the shear zone is the $L(P)_2$ stretching lineation. This is commonly developed in phyllonites and the muscovite-quartz schist slices within the shear zone. Stereoplots of these lineations show them to have a consistent plunge towards the north-northeast, varying between 65° to 70° (Fig. 4.3).

4.3.2.5.3 Displacement

(i) Macrostructures

The Ratelfontein suite shows a left-lateral displacement of 12 km along the Steenbok Shear (Annexure 1). There is no doubt therefore that strike-slip movement accounts for the major displacement along this shear. However, $L(P)_2$ plunges steeply in the near vertical shear zone indicating that at least some of the total net slip must have come from near vertical uplift of crustal blocks along the steeply-dipping shear zone (Fig. 4.17).

In the calculation of displacement along this shear zone an average value of 360° for the fault strike is used (Fig. 4.18(a)). Average strike and dip values of strata adjacent to the fault result in the following displacements being measured from piercing points on the fault plane: Net slip 12.4 km; strike slip component 8.8 km and dip slip component 6.3 km (Fig. 4.18(b) and (c)).

(ii) Mesostructures

The stretching lineation ($L(P)_2$) and fold axes of various styles reveal that the main displacement directions include an uplift of the western block along the lineation direction, preceded by a major left-lateral displacement.

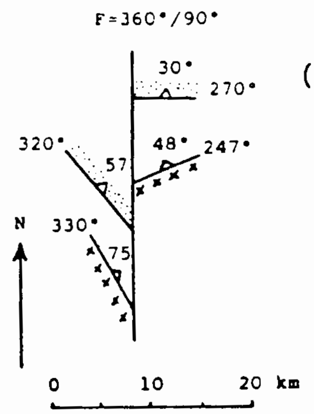
A crenulation cleavage axial planar to the isoclinal/tight folds within the shear, developed in response to the last easterly-directed tectonic event. Small crinkle and kink folds also developed during this tectonism.

(iii) Microstructures

The following microstructures were the most reliable criteria for deciphering movement directions along the shear: asymmetric shape to quartz ribbons, asymmetric "fish" micas (muscovite), cross-cutting S-surfaces, and quartz grains recrystallized with their long axes oblique to $S(P)_2$. These criteria are essentially those proposed by Simpson and Schmid (1983) (cf. Fig. 4.6).

Quartz ribbons are common within the Steenbok Shear as well as most of the other shear zones. Their asymmetric shape, as well as that of muscovite "fishes" within the fabric of the sheared rocks, show that the western block moved up relative to the eastern block (see Plate 13 and Fig. 4.19(a)). This was confirmed by an imbricate pattern of elongate quartz grains aligned approximately 30° to the foliation, in the extreme west of the shear zone. These grains recrystallized with their long axes oblique to the foliation, during the upward movement of the western block (Fig. 4.19(b).

Towards the southern end of the Steenbok Shear cross-cutting relationships seen with respect to S(P) foliations reveal several stages of development (Fig. 4.20(a) and (b)). The foliations here are enhanced chiefly by the strong alignment of elongate laths and needles of muscovite, but also by the asymmetric shape to some of the quartz ribbons (Fig. 4.20(b)). $S(P)_2$, shown in the centre of Fig. 4.20(a), lies at an angle of some 30° to the main foliation, $S(P)_3$, within this rock. $S(P)_4$ lies at an angle to $S(P)_3$, clearly disrupts it, and has therefore developed slightly later.



(a) Plan of Steenbok Shear showing displacement of Ratelfontein suite (dots) and Sabieboomrante adamellite (crosses).

(b) Stereogram data.

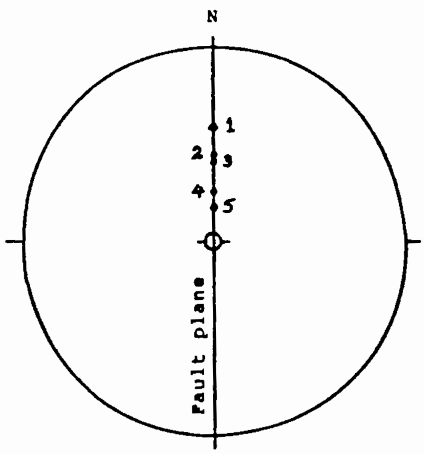
1 = pitch of Ratelfontein suite east of fault (30°)

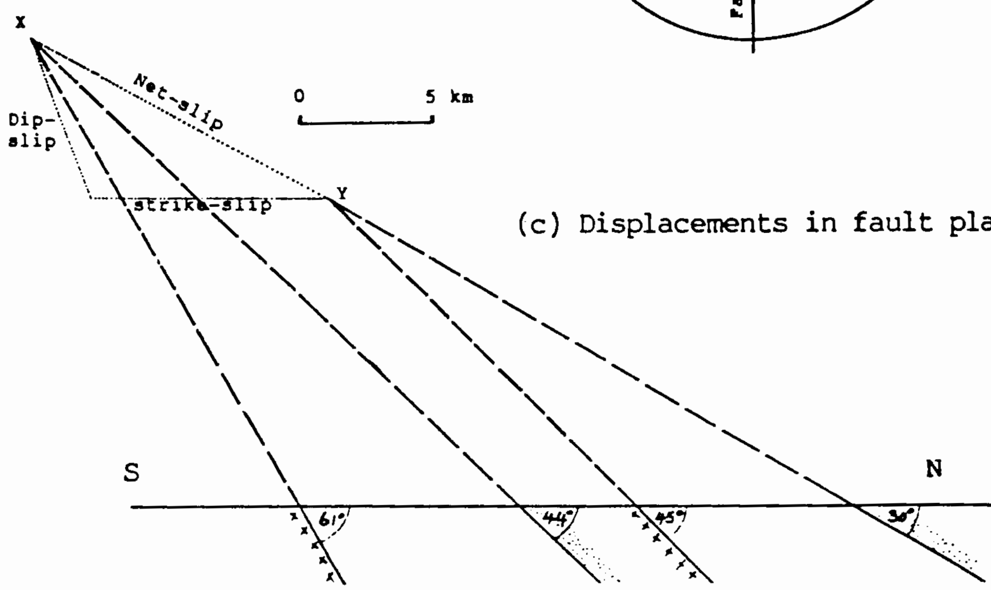
2 = pitch of Ratelfontein suite west of fault (44°)

3 = pitch of adamellite contact east of fault (45°)

4 = pitch of adamellite contact west of fault (61°)

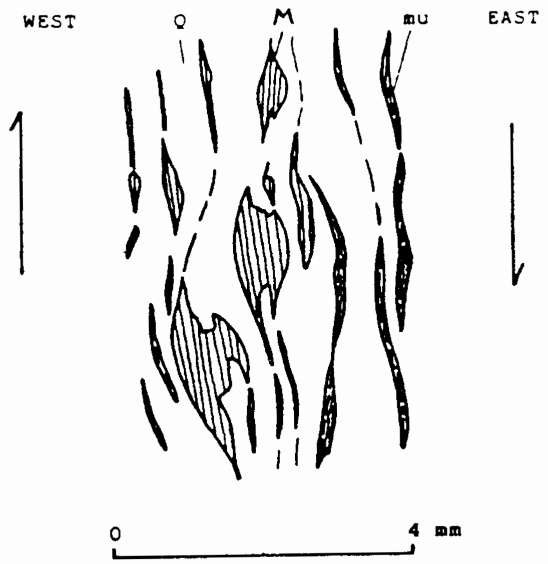
5 = pitch of $L(P)_2$ in fault (70°)



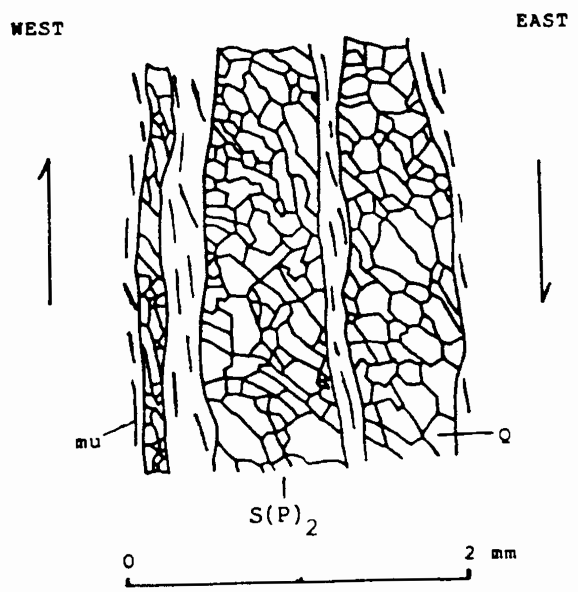


(c) Displacements in fault plane.

Fig. 4.18 Data and methods used to calculate displacements on Steenbok Shear.

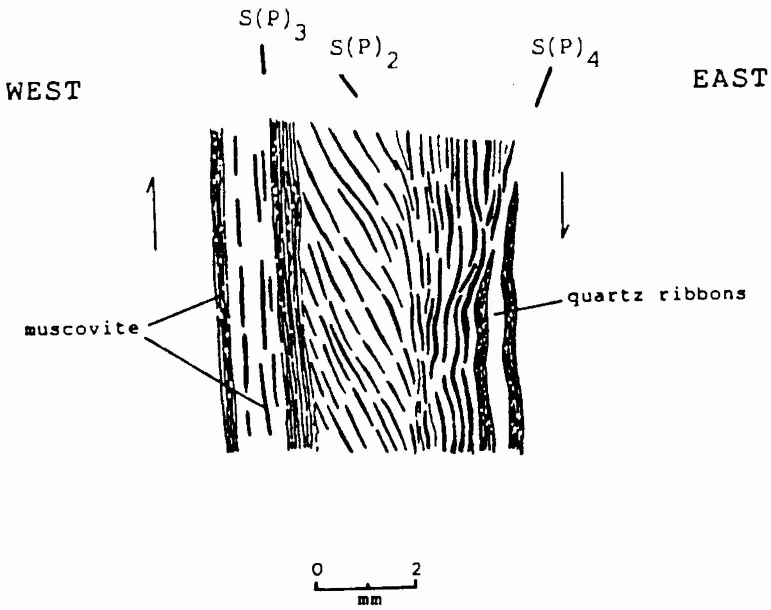


(a) Muscovite "fishes" (M) separated by quartz matrix (Q) and muscovite folia (mu) indicate a dextral sense of movement. Sketched from a thin section.

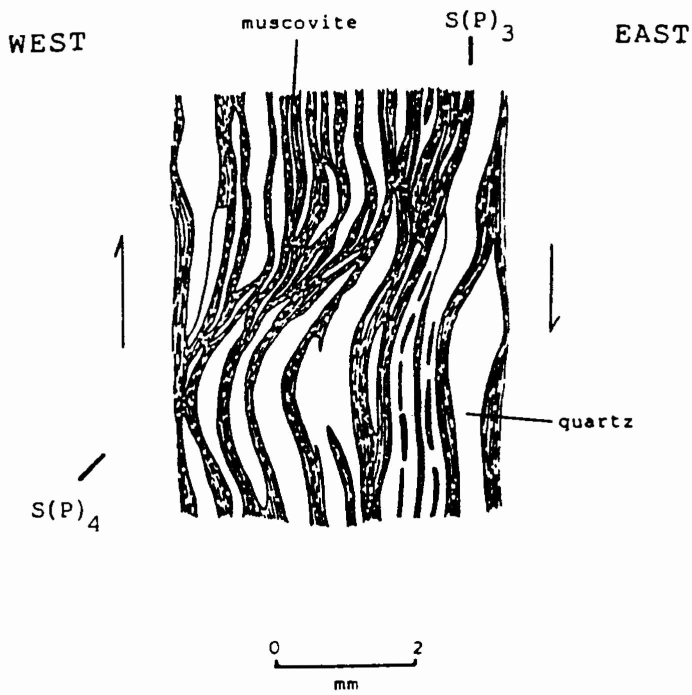


(b) Recrystallized quartz grains (Q) within ribbons oriented at an oblique angle to the foliation ($S(P)_2$) indicate a dextral sense of movement. mu = muscovite. Sketched from a thin section.

Fig. 4.19 Microstructures associated with the Steenbok Shear.



(a) Three cleavages - $S(P)_2$, $S(P)_3$, $S(P)_4$ - in the Steenbok Shear. A dextral sense of shear is indicated from the intersecting cleavage relationships. Sketched from a thin section.



(b) Development of youngest cleavage, $S(P)_4$. A dextral sense of shear is indicated from the asymmetric shape of quartz ribbons. Sketched from a thin section.

Fig. 4.20 Intersecting cleavages, Steenbok Shear.

4.3.2.5.4 Younger shears related to the Steenbok Shear

Younger shear zones, the majority less than one metre wide, are sporadically developed throughout the area. They manifest best in the northern part where they have developed a dominant north-striking fabric within the Windvlakte volcanics (see Subdomain 3.1, Domain 3, Annexure 3). They appear to splay from the Steenbok Shear and assume a dominantly northerly trend with a steep dip. Where they traverse the Vioolsdrif granodiorite they contain well developed biotite selvedges, usually several centimetres long, which form a lineation plunging northwards at steep to moderate angles (see Structural Subdomain 3.1, and Fig. 4.3).

These shears appear to be slightly younger than the major shears in the area and were probably developed during the reactivation stage that produced the $S(P)_3$ and $S(P)_4$ structures of the Steenbok Shear.

4.3.2.6 Tierkloof Shear

The Tierkloof Shear, located some 5 km west of the Steenbok Shear, varies in strike from slightly west of north to slightly east of north, with a predominantly steep westward dip. It is up to 30 m in width, and is made up of a quartz-muscovite schist. Bands of whitish massive quartzite have been incorporated within this shear zone, but have been flattened parallel to the prominent $S(P)_1$ direction (grid reference J9, Annexure 1).

The apparent sense of movement is right-lateral, opposed to that indicated by most of the other shear zones. However, the disposition of strata is best explained by a high angle reverse fault, the maximum principal stress direction oriented north-west to south-east. The western block has moved up relative to the eastern block, parallel to the $L(P)_2$ lineation direction (Fig. 4.21).

Treating this shear zone as a simple reverse fault, with an average strike of 360° and a dip of 90°, the following values of displacement of strata either side of the fault are measured: Net slip 16.8 km, strike-slip component 7 km and dip-slip component 13.5 km (Fig. 4.22). In addition to showing dip-slip and strike-slip components, the fault has rotational characteristics, the western block having been uplifted and rotated clockwise relative to the eastern block.

4.3.2.7 Kromnek Shear

The Kromnek Shear is very prominent, averaging 30 m wide, and is located some 5 km west of the Tierkloof Shear. It strikes 200° on

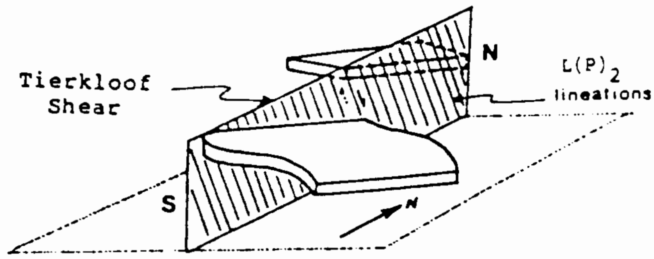
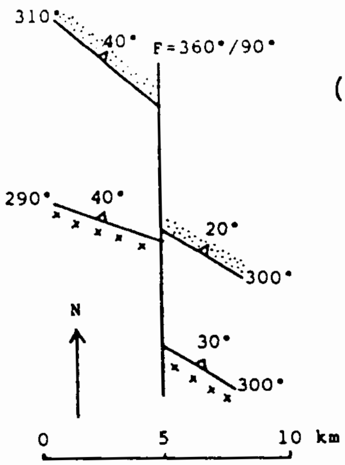


Fig. 4.21 Schematic diagram showing offset position of Ratelfontein suite strata along the Tierkloof Shear. Arrows in shear zone indicate sense of movement parallel to $L(P)_2$(strata on west side of shear have moved upwards relative to those on the east).

average, dips 85° westwards. $L(P)_2$ lineations within the shear plane plunge steeply to the north-west with an average orientation of 355°/76°.

The shear zone is composed of a quartz-muscovite schist. It shows a granoblastic-polygonal groundmass of quartz grains, separating larger quartz ribbons. $S(P)_2$ is enhanced by lepidoblastic alignment of needle like laths of muscovite (see Fig. 4.24). The rock is also characterized by muscovite porphyroblasts exhibiting undulose extinction, and asymmetric fish shapes. In places a sericitic foliation is present which clearly truncates the muscovite porphyroblasts, and enhances the asymmetric shape to quartz ribbons and augen. Microscopic evidence therefore reveals that the western block has moved up relative to the eastern block, during the last tectonic episode.

On a macro-scale a minimum strike-slip displacement of 7-9 km is indicated from the surface offset of the Ratelfontein suite (Annexure 1). The average dip of the latter west of the shear zone is steeper than that to the east, indicating a rotational component, displacement along the fault being variable (Fig. 4.23). The centre of rotation lies just to the south of the southern limit to the area. A paucity of outcrop in the south-western part of the area necessitated using structural and lithological data from farther south of the present area (Joubert, 1971; sheet 2971) in order to calculate displacements along this shear (Fig. 4.25). A dip-slip of 3.1 km is measured on the fault plane section, a rotation angle of some 12° satisfying the geometrical configuration of strata either side of it. The downthrown side is on the west, opposite to most other shears in the area.



(a) Plan of Tierkloof Shear showing displacement of Ratelfontein suite (dots) and Sabieboomrante adamellite (crosses).

(b) Stereogram data.

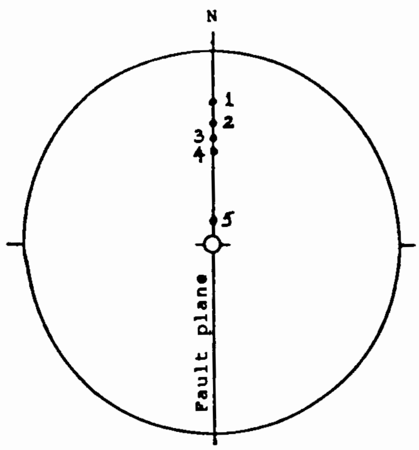
1 = pitch of Ratelfontein suite east of fault (17°)

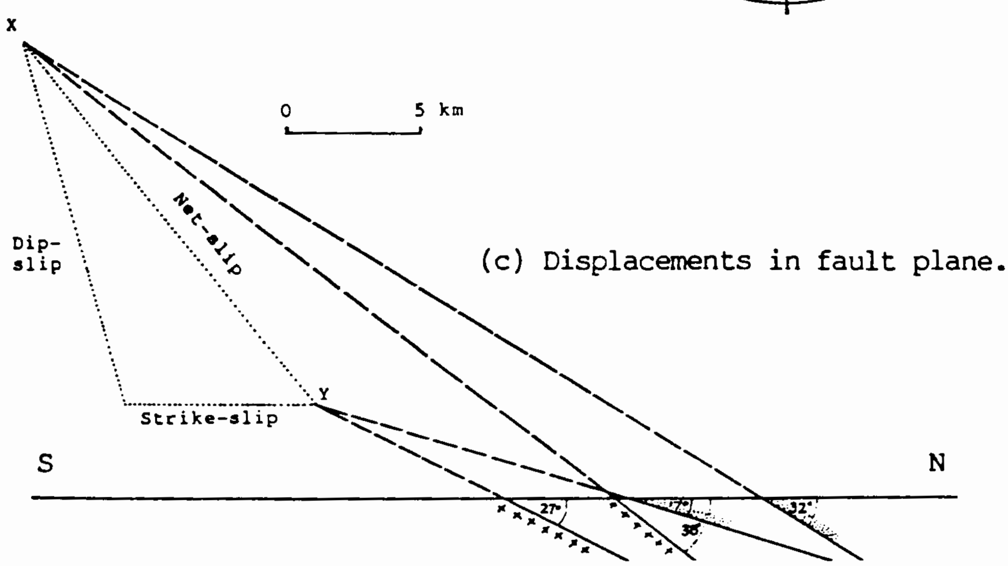
2 = pitch of adamellite contact east of fault (27°)

3 = pitch of Ratelfontein suite west of fault (32°)

4 = pitch of adamellite contact west of fault (38°)

5 = pitch of $L(P)_2$ in fault (75°)





(c) Displacements in fault plane.

Fig. 4.22 Data and methods used to calculate displacements on Tierkloof Shear.

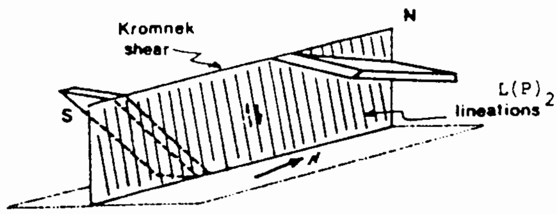


Fig. 4.23

Schematic diagram showing present left-lateral offset position of Ratelfontein suite strata along Kromnek Shear. Strata have been rotated into different orientations either side of the fault. Arrows in shear zone indicate movement direction parallel to $L(P)_2$ along which strata on the west side of the fault moved upwards, after the rotational episode.

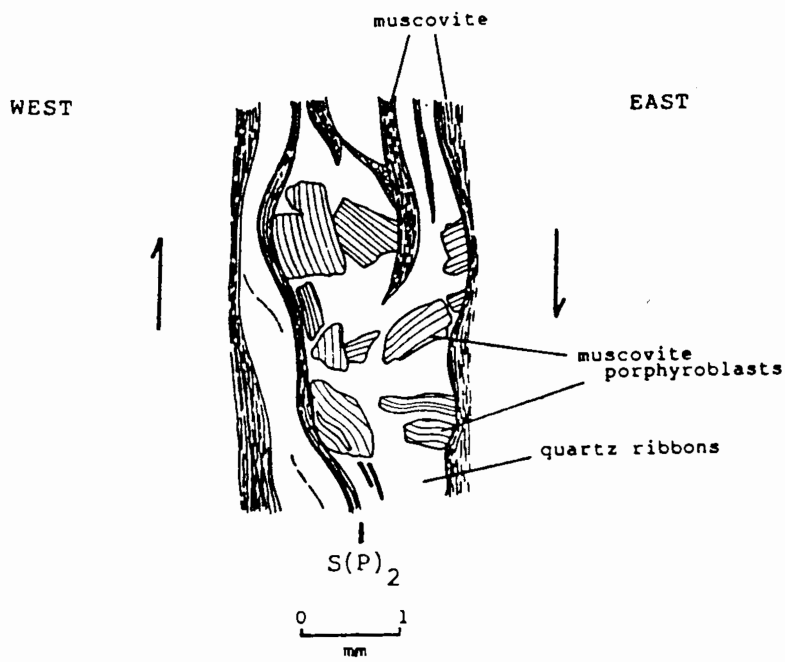
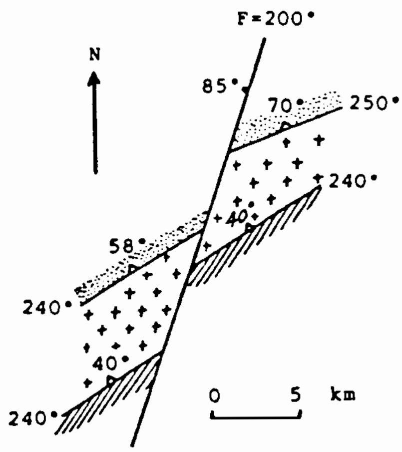


Fig. 4.24

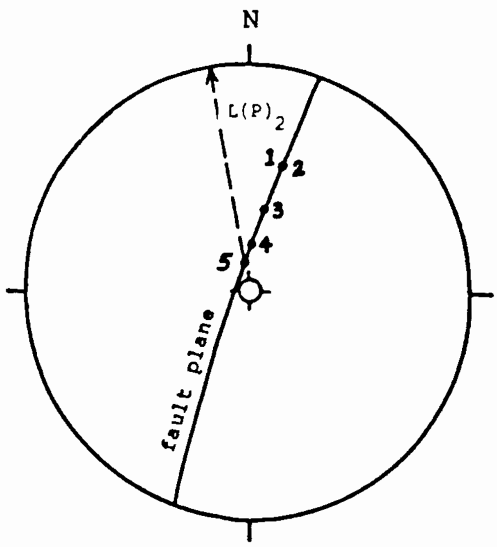
Microfabric of quartz muscovite schist, Kromnek Shear. Orientation of muscovite porphyroblasts and shape of quartz ribbons show dextral sense of shear during last tectonic episode. Sketched from a thin section.



(a) Plan showing displacement of Ratelfontein suite (dots), Kouefontein granite gneiss (crosses) and Chabiesies suite (cross-hatching) along the Kromnek Shear.

(b) Stereogram data.

1 & 2 = pitch of Chabiesies suite top contact west and east of fault (30°)
3 = pitch of Ratelfontein suite (base) west of fault (50°)
4 = pitch of Ratelfontein suite (base) east of fault (68°)
5 = pitch of $L(P)_2$ in fault (76°)



(c) Rotational displacement in fault plane.

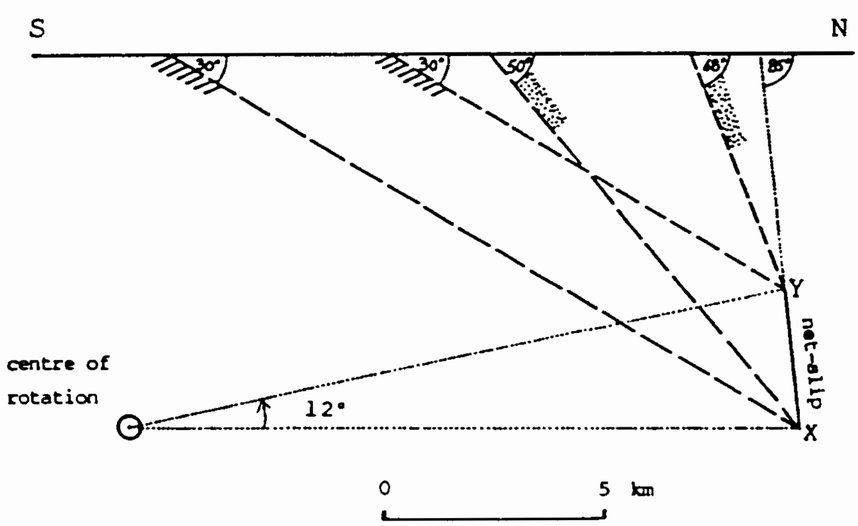


Fig. 4.25 Data and methods used to calculate displacement on Kromnek Shear.

The last major movement, as shown by the arrows within the shear zone is a reverse (thrust?) motion which took place after the main rotational movement. This sense of movement is borne out by microscopic evidence (Fig. 4.24). However, there must have been a strike-slip component involved prior to the rotational tectonic episode in order to offset the Ratelfontein suite so far south on the western side of the shear zone (Annexure 1). If a strike-slip movement did not initially take place then the "reverse" faulting episode would have offset the Ratelfontein suite northwards on the western side of the fault.

4.3.2.8 Stinkfontein Contact Shears

For some 2 km eastwards from the Stinkfontein contact shear zones are closely spaced, so that $S(P)_2$ is here the dominant fabric element. It has an average strike of 193°, and dips 86° westwards (Subdomain 1.1, Domain 1, Annexure 3). $L(P)_2$ here also plunges steeply to the north-west (Fig. 4.3).

Shearing along the contact was documented earlier by Rogers (1915) and De Villiers and Söhnge (1959). It took place over an extended period, corresponding to the "Devils Castle event" (Ritter, 1980). The Gariep unconformity is said to cut across this foliation nearer Eksteenfontein (op. cit., p. 136), but in the present area the main tectonic cleavage in Stinkfontein rocks is continuous with $S(P)_2$ in the pre-Gariep basement. This observation is significant in that it places the development of $S(P)_2$ (during Pan-African deformation) as post-Stinkfontein Formation in age.

4.3.2.9 Summary statement

Analysis of displacement along the major Pan African shear zones shows that the intervening crustal blocks in the area have been displaced and rotated relative to each other. This rotation took place in response to south-easterly directed compression which developed high-angle faults with a variable southerly strike-slip component along some of the shears. Calculated displacements are summarized in Table 4.2. These figures show only total displacements and do not reflect the history of movement along each shear. Each fault block is uplifted relative to the block on its eastern side, the only two exceptions being the Witbakensberg and Kromnek Shears, where the western blocks are downthrown.

The abrupt western limit to the Nama Group sediments against the Steenbok Shear is explained by uplift of the crustal blocks on its western side in post-Nama times (see Section 4.5.3.5), followed by erosion to the present topographic level. This addresses the anomaly posed in section 2.7.2.4.

TABLE 4.2 Calculated displacements on Pan-African shears			
Shear	Displacement in km		
	net-slip	strike-slip	dip-slip
Witbakensberg	0,3	---	0,3
Riethoek	7,5	0,9	7,3
Steenbok	12,4	8,8	6,3
Tierkloof	16,8	7,0	13,5
Kromnek	3,1	7,0 - 9,0?	3,1

4.3.3 Anomalous features associated with Pan-African shears

4.3.3.1 Distribution of supracrustal rocks

There is a gross difference in width of outcrop of the Groothoek suite along strike (Section 2.7.2.1), especially between the Tierkloof and Steenbok Shears, where these predominantly meta-psammitic rocks outcrop over a broader area than to the west or east (central area of Annexure 1). The Groothoek suite has an assumed three-dimensional wedge geometry, the strata thickening towards the north and pinching out westwards (cf. Fig. 3.11). Subsequent disruption along the northerly striking Pan-African Shears raised the western crustal blocks relative to those on their easterly side. Groothoek strata thickening down dip west of the Steenbok Shear will therefore have a broader outcrop than equivalent strata at the same present elevation on the east side (Fig. 4.26). A slight southwards rotation of the block to the west of the Steenbok Shear may well have assisted in juxtaposing a greater width of outcrop of the Groothoek suite. The same strata west of Tierkloof Shear have a narrow outcrop because they wedge out rapidly westwards, just beyond this shear zone (cf. Fig. 3.11).

Between Steenbok and Tierkloof Shears, only a very small outcrop of rock types belonging to the Groenrivier suite overlies the Groothoek suite (cf. Fig. 3.1, and Annexure 1). Where these rocks are exposed they are truncated by the Vioolsdrif granodiorite (grid reference K5, Annexure 1). East of Steenbok Shear and west of Tierkloof Shear, the Vioolsdrif granodiorite is in contact with the stratigraphically higher Groenrivier suite. To explain this

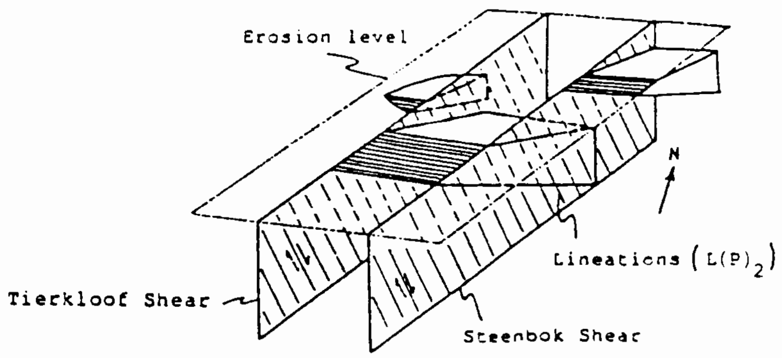


Fig. 4.26 Schematic drawing showing offset position and broad outcrop pattern of Groothoek suite strata between Steenbok and Tierkloof Shears. Note lineations indicate movement direction after a proposed left-lateral displacement along the Steenbok Shear had taken place.

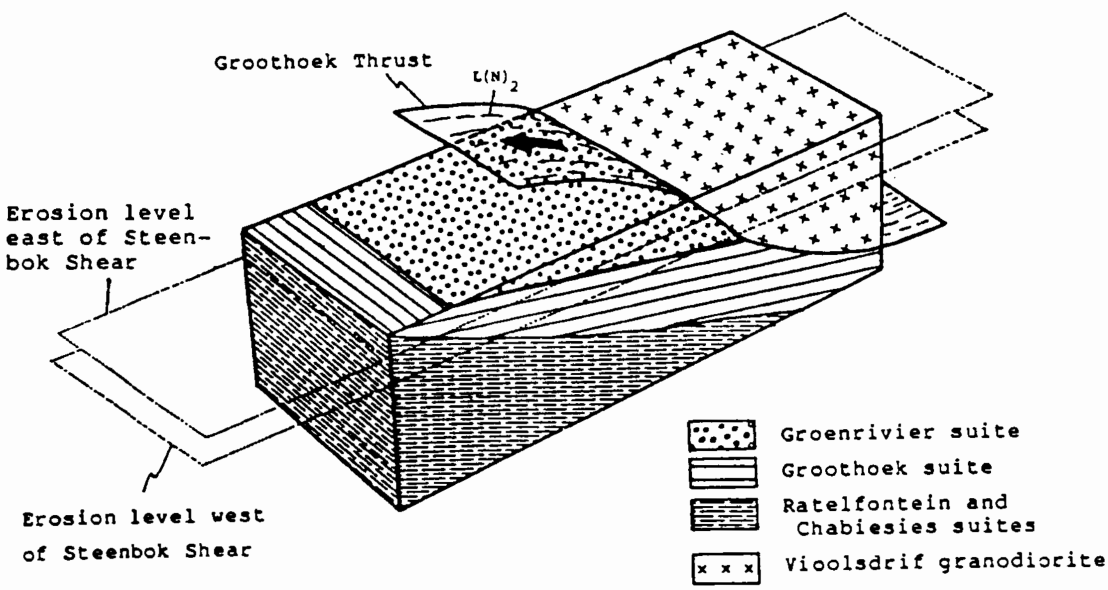


Fig. 4.27 Schematic diagram illustrating variation in outcrop width of strata at different erosional levels, due to transgressive nature of Vioolsdrif intrusion with depth, south of Groothoek Thrust. Note that Groenrivier suite has broader width of outcrop at erosion level west of Steenbok Shear than to the east.

configuration it is suggested that Vioolsdrif granodiorite probably has a transgressive contact and steepens in depth, tectonically slicing across the strata, so that erosion of fault blocks to different levels exposes this transgressive relationship (Fig. 4.27).

The Windvlakte metavolcanics (upper unit) have a restricted geographical distribution between Tierkloof and Steenbok Shears, and west of Kromnek Shear (Section 2.7.2.2). This may be due to a large left-lateral component along the Steenbok and Kromnek Shears. This has indeed taken place along the Steenbok Shear in post Nama times (see section 4.3.2.5). In the case of the Kromnek Shear a strike-slip movement probably occurred prior to development of the high-angle thrusting (see section 4.3.3.4).

4.3.3.2 Change in strike of shears

Although the Pan-African shear zones are diagramatically depicted as perfect planar structures (cf. Fig. 4.7), most do not have a constant orientation. Their strike varies from slightly west to slightly east of north (e.g., the Steenbok Shear), dips being predominantly westward, ranging from 65° to vertical.

A consistent pattern to the strike variation is apparent along the southernmost section of nearly every shear zone (cf. Fig. 3.1 and Annexure 1). Here the strike changes either to due south, or swings south-southeasterly. The change occurs mostly where the shear zone traverses granitic-type lithologies, e.g., Sabieboomrante adamellite gneiss. These massive rock types probably acted as resistant buttresses during development of the major shear zones, probably due to competency contrasts with the supracrustal lithologies, therefore causing deflections of shear zone strike.

In Fig. 4.30, the restored position of Pan-African shear zones is shown at the onset of Late Proterozoic deformation. The most pronounced deflection occurs along the southern Kouefontein Shear. Here the characteristic en echelon pattern swings towards the south-east, at the contact between Kouefontein granite gneiss and Sabieboomrante adamellite gneiss. Although the outcrop trace of the individual en echelon shears is variable, they probably join up in depth with a ductile shear of even greater south-eastwards deflection, approximately parallel to the top contact of the Sabieboomrante adamellite gneiss (see Fig. 4.28).

4.3.3.3 Steepening dip of strata adjacent to shear zones

Strata adjacent to some shear zones steepen up significantly and may dip vertically within the shear. Steepening up of the Kouefontein synform adjacent to the Kouefontein Shear (section

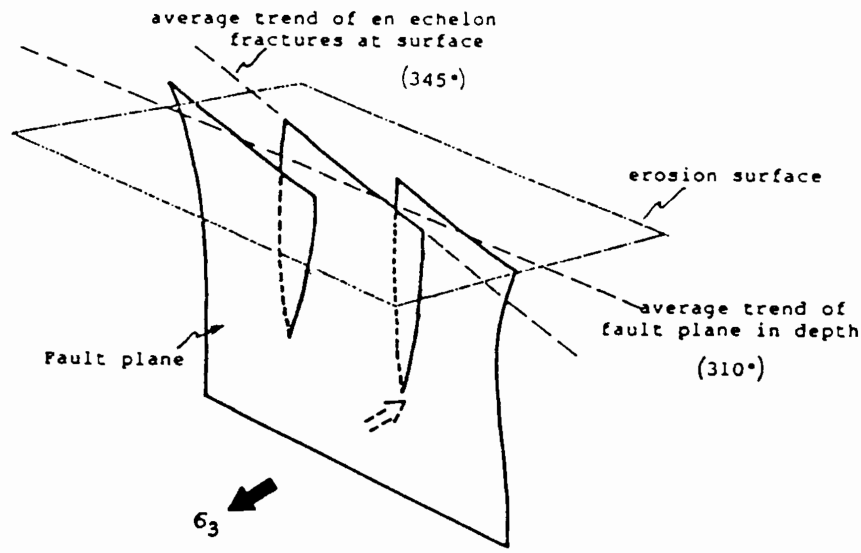


Fig. 4.28 Schematic drawing showing the relationship of Kouefontein en echelon shears on surface to possible fault plane at depth (modified after Park, 1983, Fig. 11.4).

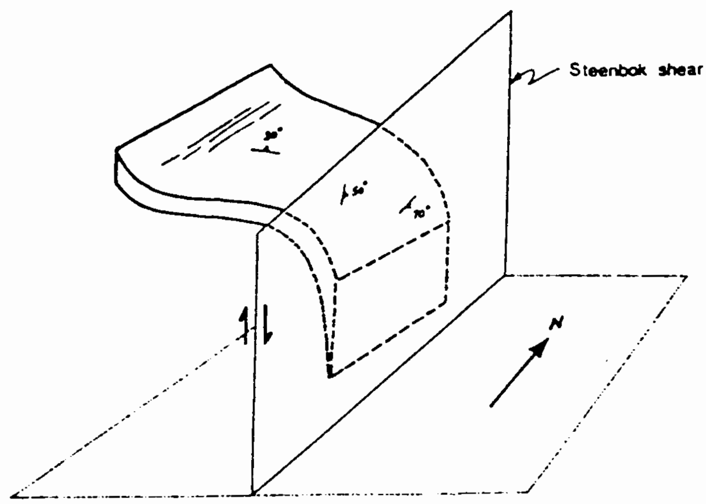


Fig. 4.29 Schematic diagram showing steepening dip of strata immediately west of Steenbok Shear. Arrows in shear zone indicate sense of movement which uplifted the western block relative to the east block, during the Gariepian thrusting episode.

4.3.2.4) involved a rotational movement of the block to the west, dragging the Ratelfontein quartzites into the shear zone.

Groothoek strata immediately west of Steenbok Shear dip at ever larger angles as the shear is approached (Fig. 4.29). This could be interpreted either as a reverse drag feature, similar to those described in extensional environments (Hamblin, 1965; Proffett, 1977; Wernicke and Burchfiel, 1982), or simply as a drag effect related to uplifting of the western crustal block. The latter interpretation is preferable because microstructures confirm the appropriate sense of movement in the Steenbok Shear (cf. Figs. 4.19 and 4.20).

4.3.3.4 Anti-clockwise rotation of $S(N)_2$

$S(N)_2$ in structural Subdomain 5.3 (Domain 5, Annexure 3) immediately east of the central portion of the Steenbok Shear strikes towards the north-east, with a north-westerly dip (247°/48°NW). Its regional orientation here diverges some 60° anti-clockwise from the average strike farther east, implying that the foliation immediately adjacent to this eastern portion of the Steenbok Shear has been passively rotated in an anti-clockwise sense.

The rotation of $S(N)_2$ here is ascribed to the left-lateral movement along the Steenbok Shear during the Late-Proterozoic/ Early-Palaeozoic history of the area. I propose that similar orientations of $S(N)_2$ in the crustal blocks farther west are the result of rotation resulting from left-lateral movement along shears in the western part of the study area (see Subdomains 2.1 and 2.2, Domain 2; and Subdomain 3.2, Domain 3, Annexure 3). Some are minor shears within crustal blocks west of Tierkloof Shear, but are not actually shown on Annexures 1 and 3. In the west, adjacent to the Stinkfontein Formation unconformity, the orientation of some strata deviates significantly from the regional north-westerly strike. The Ratelfontein suite, for example, has a variable south-westerly strike (grid reference C9, Annexure 1), interpreted to be the result of left-lateral shearing along the Stinkfontein Contact Shears.

4.3.3.5 Interference folding, Kouefontein Synform

The Kouefontein Synform represents a gentle fold plunging towards the east-northeast (cf. Fig. 4.14). The main structure in $S(N)_2$ trends towards 043°, and plunges at 22° (Subdomain 5.2, Domain 5, Annexure 3). The axial trend of this synform is at an angle of approximately 30° to the strike of the Kouefontein and Steenbok Shears.

Axes of open folds in this area plunge into the north-western and south-eastern quadrants, indicating that they belong to an earlier generation which has subsequently been folded about the north-easterly plunging gentle fold (the Kouefontein Synform).

The main synform seems to have developed during left-lateral movement along the Steenbok Shear. The Kouefontein synform structure can thus be related to numerous examples of folds which have developed synchronously with strike-slip faulting (De Sitter, 1956, p. 159; Moody and Hill, 1956; Harding, 1974; Joubert, 1974a).

4.3.3.6 Summary Statement

A number of anomalous features of lithological distribution and structure in the area can be explained as the product of Late-Proterozoic/ Early-Palaeozoic (Pan-African) tectonism within the foreland margin of the Gariep Belt.

Initial left-lateral strike-slip faulting along the Steenbok Shear and Kromnek Shears resulted in the offsetting of strata adjacent to the shears. Metavolcanics of the Windvlakte suite owe their present rather restricted geographical distribution to this initial strike-slip movement along the Steenbok and Kromnek Shears. During this movement strata of the Ratelfontein suite were displaced and oriented towards a south-westerly strike in the area west of Kromnek Shear.

Subsequent tectonic uplift and rotation along these shears gave rise to other anomalous features, e.g. the change in outcrop width of certain metasedimentary/ metavolcanic units. This is explained by uplifting of the crustal block west of the Steenbok Shear, and subsequent erosional unroofing of a northward dipping Groothoek suite metasedimentary wedge.

Granitic massifs within the Bushmanland Subprovince apparently acted as resistant physical buttresses during south-easterly directed Pan-African tectonism along the coastal belt. This effected characteristic south-southwest to south-southeast change in strike of the major shear zones along their southern outcrop traces.

The steepening dip of strata observed on the immediate western side of some of the shears is interpreted as the result of drag during upward movement of crustal blocks, during Pan-African tectonism in this area.

The Kouefontein synform is interpreted as forming in relation to left-lateral movement along the Steenbok and Kouefontein Shears.

4.4 A PALINSPASTIC RECONSTRUCTION

Restoration of sections and plans was carried out to produce a palinspastic map which shows the geological configuration just prior to 900 Ma (Fig. 4.30). This was constructed by reversing the net-slip of each crustal block, in turn, along the relevant north-north-east trending shear zone, and adjusting for any rotational component, where applicable. The continuity of reliable marker units such as the Ratelfontein suite proved to be valuable in this task.

This simplified palinspastic reconstruction confirms three significant features.

(i) The continuity of lithologies is clearly indicated, giving a better idea of the three-dimensional distribution of supracrustal rocks in particular (cf. section 3.1.6). The correlation and stratigraphic position of certain plutonic intrusive rocks is clarified (cf. section 3.2.3). The Sabieboomrante adamellite gneiss and the overlying Kouefontein granite gneiss are very large conformable sheet-like bodies within the north-eastward dipping Chabiesies suite of the Een Riet Subgroup, and both extend across nearly the entire area. The plutonic rocks of the Richtersveld Subprovince are relatively homogeneous, but are lithologically more variable in the localized north-west corner.

(ii) Pre- Pan-African structures, particularly thrust zones, are evidently continuous across most of the area. The projection of these thrust zones west of Steenbok Shear is in some cases subjective, especially in regions showing little outcrop.

(iii) Where the north-north-east trending Pan-African shear zones cut across coarse-grained granitic type rocks their strike changes towards due south and south-east. These granitic rocks are interpreted as forming physical buttresses during development of the shears because of their massive nature. The shears either deflect around them, or, in some cases, develop en echelon patterns at the contact zones.

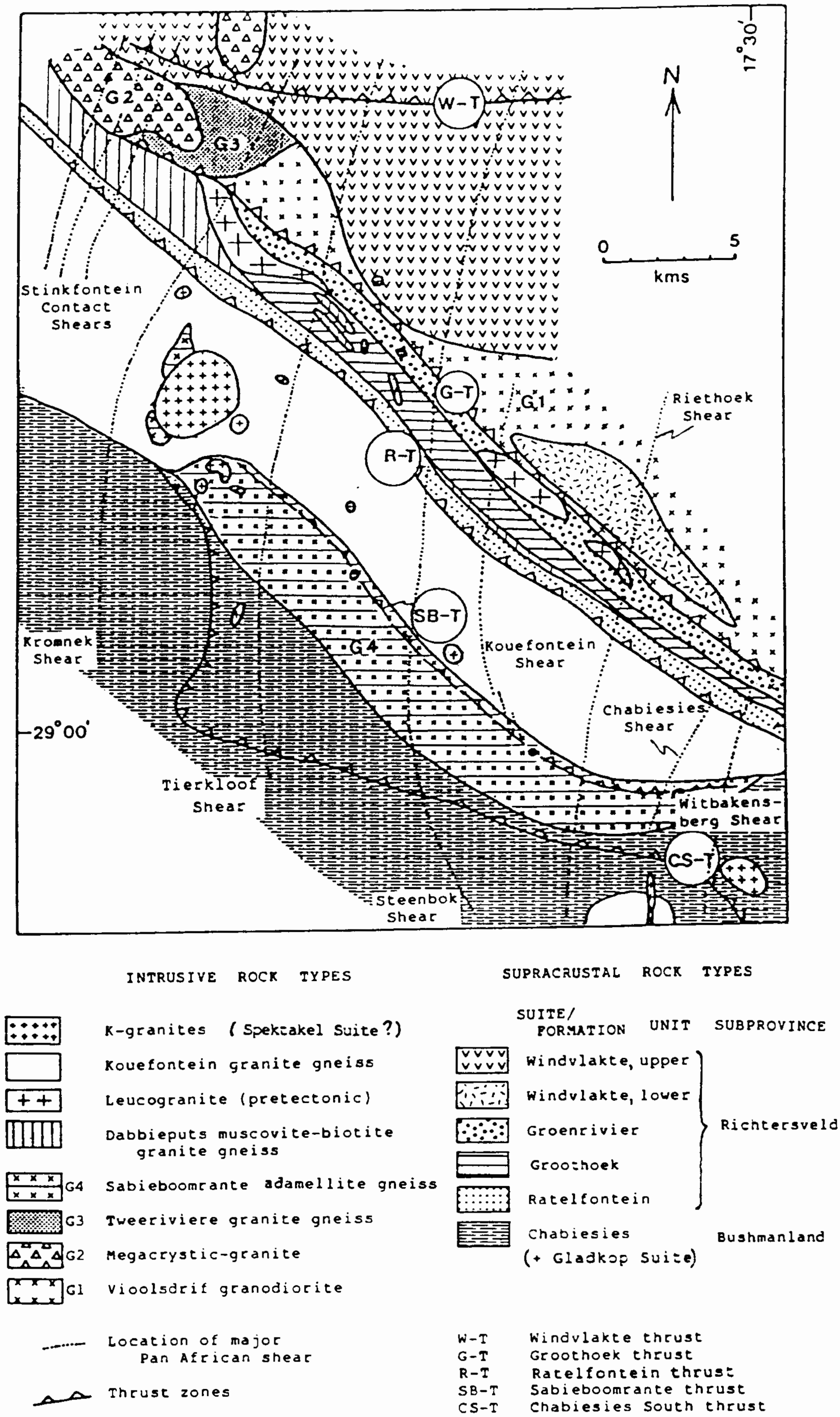


Fig. 4.30 Palinspastic reconstruction to Late Proterozoic times.

4.5 NAMAQUA SHEAR ZONES AND RELATED STRUCTURES

Thrust zones parallel the regional Namaqua ($S(N)_2$) foliation (Fig. 4.31). The numbering sequence reads from north to south, the eastern sector containing most of the thrust zones. Thrust zone strike continuity has been offset along Late- Proterozoic/ Early Palaeozoic shear zones, but correlation across the area is nevertheless possible from a palinspastic reconstruction to pre-Pan-African times (cf. Fig. 4.30).

4.5.1 Windvlakte Thrust

4.5.1.1 Locality just west of Steenbok Shear

(i) Field appearance

In the north-centre of the area a prominent mylonite zone averaging 2 m wide occurs in metavolcanics of the Windvlakte suite (grid reference L1, Annexure 1). It does not continue into adjacent Vioolsdrif granodiorite and is therefore older than the latter. In outcrop it has a weathered, fissile appearance, the mylonitic foliation striking 265° and dipping 45° northwards.

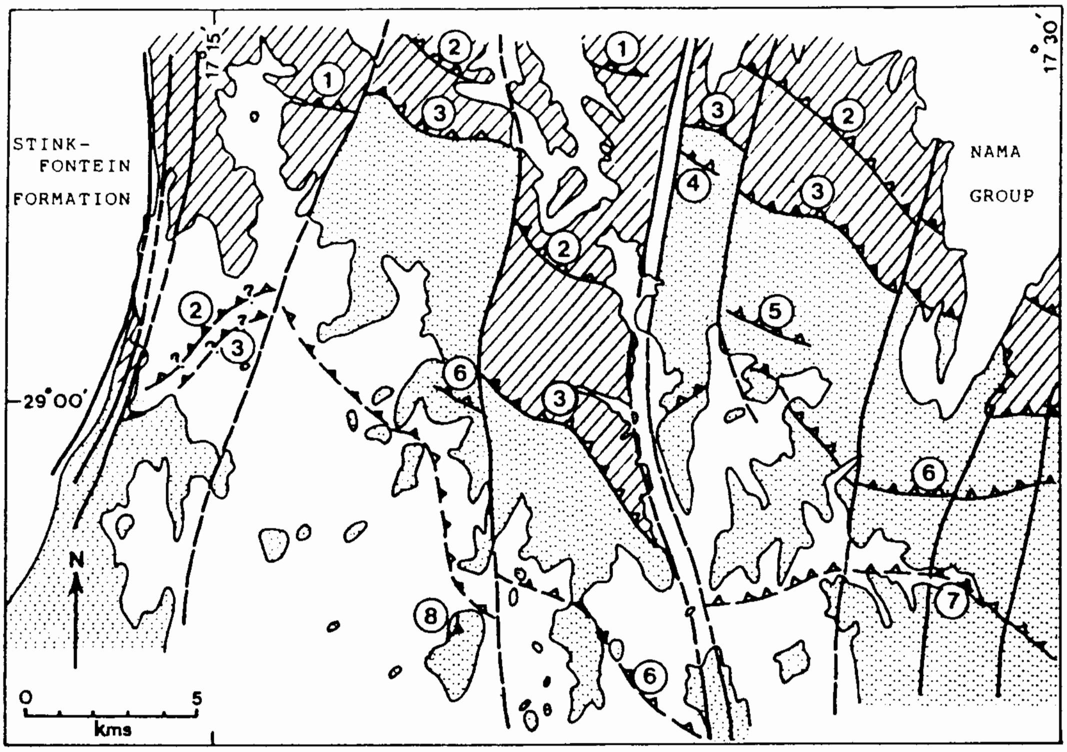
A poorly developed lineation ($L(R)_1$, Table 4.1) plunges down-dip (005°/45°). Towards the top of the deformation zone the mylonitic foliation gives way to an anastomosing foliation which is less pronounced upwards through the zone, and gradually disappears into the hangingwall.

(ii) Petrography

The mylonite has a well developed quartz-ribbon texture with clasts of plagioclase and quartz set in a fine-grained groundmass of mainly biotite, epidote and quartz (Fig. 4.32, and Plate 14). The mylonitic foliation (Sm) is outlined by parallel alignment of biotite and small epidote grains. The rock is traversed by a subsequent cleavage (Sc) at an angle of about 30° to Sm.

Quartz ribbons are of variable width, averaging 0,4 mm and are strung out continuously across the entire slide. They are thinned in places, especially where Sc transects them. The grains are all recrystallized with high-angle boundaries, indicating that recovery is complete. The ribbon boundaries are irregular where in contact with groundmass biotite and epidote. Porphyroclasts of plagioclase up to 1,5 mm in size show rigid-body rotation.

Muscovite forms thin needle-like laths 0,05 mm long, parallel to Sm. Olive-green biotite makes up about 25% of the rock, occurring as small broad laths <0,1 mm long. Both phyllosilicates outline the strongly developed foliation (Sm). Biotite is also developed across Sm at an angle of 30° thereby enhancing Sc (cf. Fig. 4.32).



Richtersveld Subprovince

Bushmanland Subprovince

Pan African Shear zone, outcrop
Pan African Shear zone, concealed
Pre-Pan African thrust zone, outcrop
Pre-Pan African thrust zone, concealed

1 Windvlakte thrust (W-T)
2 Groothoek thrust (G-T)
3 Ratelfontein thrust (R-T)
4 Kouefontein North Ramp
5 Kouefontein mylonite gneiss
6 Sabieboomrante thrust (SB-T)
7 Chabiesies South thrust (CS-T)
8 Kabies se Berg thrust

Fig. 4.31 Location of major thrust zones in study area.

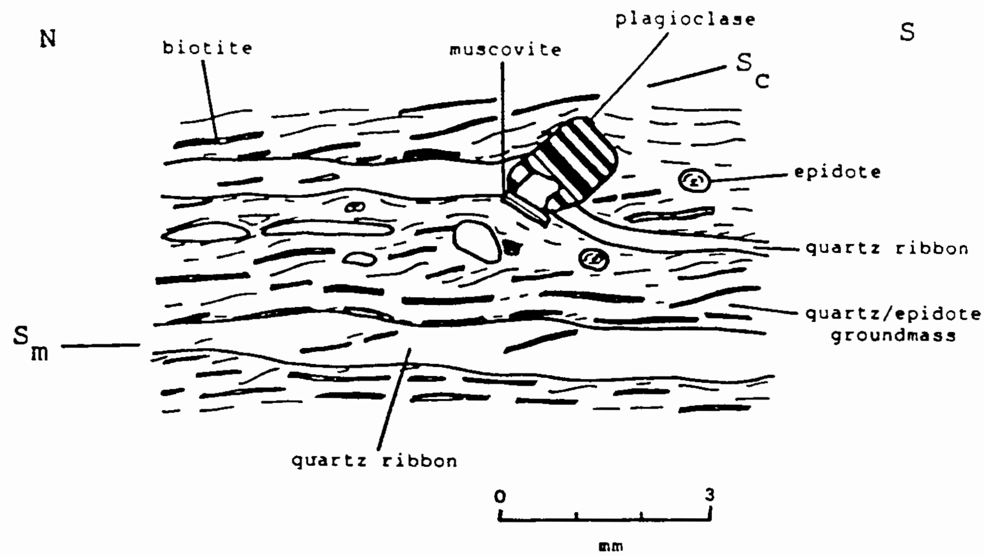


Fig. 4.32 Mylonite in Windvlakte Thrust. Note pronounced development of quartz ribbons which define the mylonitic foliation (Sm), and laths of biotite formed along a cleavage (Sc) at 30°to Sm. A dextral sense of shear is interpreted from the orientation of porphyroclasts and asymmetric shape of quartz ribbons. Drawn from a photomicrograph.

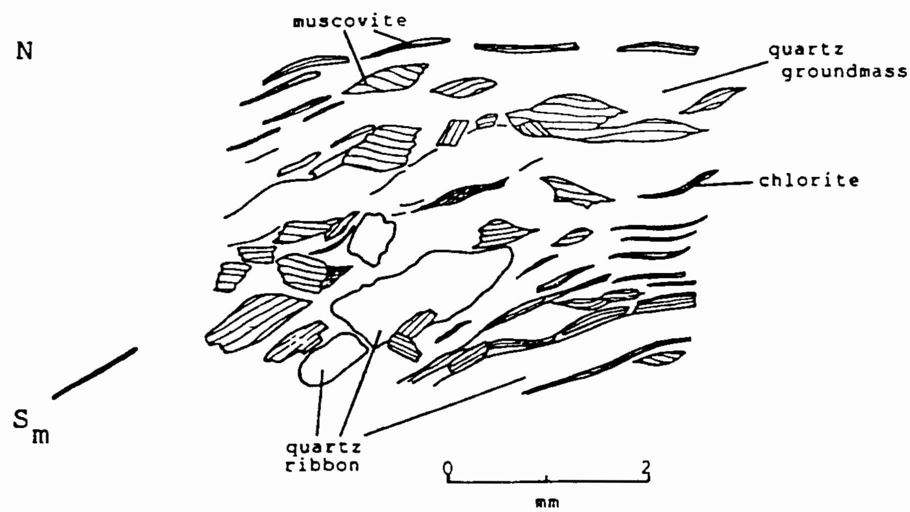


Fig. 4.33 Development of quartz augen and ribbons, and fish-shaped muscovites showing a dextral sense of shear in Windvlakte Thrust, west of Kromnek Shear.
Drawn from a photomicrograph.

Abundant small rounded epidote grains averaging 0,01 mm in size occur parallel to Sm. Occasional, randomly oriented larger grains (up to 0,2 mm) are also present.

In the central portion of the thrust zone the mylonitic fabric is deformed into small chevron folds (F(N)3), with muscovite outlining a well developed axial planar cleavage (Plate 15). The vergence of the cleavage is towards the north.

4.5.1.2 Locality just west of Kromnek Shear

(i) Field appearance

A schistose unit about 1m thick, displaying a well developed mylonitic foliation (295°/55°N) occurs in a broad zone of sheared metavolcanics of the Windvlakte suite, immediately west of Kromnek Shear (grid reference G2, Annexure 1). Within the foliation a lineation plunges steeply down-dip to the north-northeast (015°/56°), and is consistent with the regional pattern (see Subdomain 1.1a, Domain 1, Annexure 3).

(ii) Petrography

Quartz ribbons are poorly developed, recrystallized aggregates of quartz augen being more prominent (Fig. 4.33). Some inequant quartz porphyroclasts display undulose extinction within the main portion of the grains, but recrystallized embayed outlines along their margins. Aggregates of small recrystallized quartz grains with high-angle boundaries between them are transected by long slender muscovite and chlorite grains outlining their asymmetric shape.

Asymmetric, fish-shaped muscovite porphyroclasts up to 1,5 mm wide show slip along their basal (001) planes, their dip being opposite to the overall sense of shear (cf. Fig. 4.6(d)).

Long thin needles of muscovite are strongly aligned parallel to the mylonitic foliation. Magnetite forms blebs of inequant grains stretched out along the schistosity, but is also present as euhedral grains.

4.5.1.3 Interpretation

Textural features in the mylonite west of Steenbok Shear show that it developed by recrystallization of quartz into elongate ribbons, while porphyroclasts of plagioclase, and occasional quartz grains, were deformed through rigid body rotation. A right-lateral (facing east) sense of shear is interpreted from the asymmetric orientation of transected quartz ribbons. These textures developed during upper greenschist facies metamorphism similar to those described by Simpson (1985). The orientation of Sc at an acute angle to Sm suggests that it probably developed at a slightly later period, or even at the same time as Sm. In the latter case this would represent an S-C cleavage relationship (cf. Fig. 4.6(e)), the two foliations developing during the same southward directed deformational phase.

Mylonitic textures in the thrust to the west of Kromnek Shear do not display as advanced stages of mylonitization as the one just east of Steenbok Shear. Petrographic characteristics are nevertheless consistent with the deformation mechanism proposed by Lister and Snoke (1984), and confirm a north over south sense of movement.

There are two possibilities concerning the regional continuity of the Windvlakte Thrust.
(i) The two occurrences could be the same, or related thrust surfaces.
(ii) These are two different thrust zones.

The first possibility would be difficult to prove because the thrust does not continue into the adjacent Vioolsdrif granodiorite, and Pan-African tectonic overprinting in this area precludes tracing the thrust for any distance in the field. If the thrust was once continuous across the area, as depicted in Fig. 4.30, then its surface was probably composed of one, or perhaps several lateral ramps. These ramps probably broke back to higher crustal levels westwards, because of the tectonostratigraphically higher position of the thrust west of Kromnek Shear. The movement direction would have been parallel to the ramps, i.e. towards the south.

The possibility therefore exists that these two thrust occurrences are unrelated and therefore represent separate thrusts. In view of difficulties in tracing the continuity of these thrusts, as explained above, the interpretation cannot be futher verified.

4.5.2 Groothoek Thrust

4.5.2.1 Historical aspect

Thrusts in Richtersveld rocks south of the Vioolsdrif-Goodhouse area were first delineated by Duplan (1976).

Van der Merwe (1979) and Theart (1980, p. 78) proposed the location of a thrust zone within the Groothoek schists some 20 km north of Steinkopf to account for their intensely developed fabric.

In the area north of Pofadder the Wortel belt line of mafic bodies, and association with magnesian schists, show deformation features (Joubert, 1974c). This suggests that tectonic movement (possibly thrusting) may be associated with them (P. Joubert, personal communication, 1983).

The Groothoek thrust in the Namaqua geotraverse area north of Springbok, is interpreted as a major ramp structure with a displacement of 75 km to the south (Blignault et al., 1983; Van der Merwe, 1986). A number of thrusts occur north-east of Springbok, particularly in the Geselskapbank area, where Strydom (1985) positions the Groothoek Thrust between supracrustal lithologies of the Nab Subgroup and the intrusive Vioolsdrif Suite.

Hartnady (personal communication, 1986) suggests that the line of mafic bodies occurring between the Steenbok Shear and the Pofadder lineament (some 150 km) may define a major geosuture (the Groothoek Thrust) which could possibly have formed about 1500 Ma ago. Van der Merwe and Botha (1989), however, interpret the Groothoek Thrust in the area of the Namaqua geotraverse as a major ramp structure which formed under ductile conditions, at ≈1100 Ma.

In view of the fact that the ideas expressed by the aforementioned authors were gained mainly from observations made east of longitude 17°30′, an important aspect of the present survey is to trace the position of the Groothoek Thrust west of the Namaqua geotraverse. In the study area the Groothoek Thrust is exposed in the escarpment region, where characteristic features such as modification of pre-existing structures and mylonitic rocks are developed in a broad zone of ductile deformation (Booth, 1987).

4.5.2.2 Field appearance, and related characteristics

The Groothoek Thrust zone is positioned at the boundary separating Vioolsdrif Suite rocks from those of the Groenrivier suite as a broad zone of deformation (Annexure 2). There is a pronounced linear arrangement of mafic/ultramafic rocks associated with the thrust zone (Klipbok complex, cf. Plate 8).

The effects of thrusting occur over a width of at least 2 km. The thrust cannot be traced as a single surface, but rather as zones of intense shearing several hundreds of metres wide which obscure contact relationships. Pegmatites and granitic rocks along this zone further conceal the original lithological contact relationships. There is nevertheless abundant evidence to show that this is a significant zone of Mid-Proterozoic ductile deformation. West of Steenbok Shear the contact zone is covered by scree, and no associated mafic/ultramafic rocks were located during the present survey. Tectonised rocks, often mylonitic in character, occur there instead.

Location of the thrust contact west of Kromnek Shear is conjectural as it is entirely obscured by scree deposits and windblown sand. If the western tip point of the thrust is present here then the thrust may not extend as far westwards as shown on Fig. 4.31.

(i) Regional foliation

The Vioolsdrif granodiorite's southern contact is characterized by a platy foliation, becoming more pronounced closer to the contact. This feature is enhanced by the parallel arrangement of constituent minerals such as biotite, hornblende and epidote.

West of Kliphoogte the granodiorite acquires a mylonitic texture with development of biotite and epidote folia on contact with lithologies of the Groothoek suite.

East of the Kouefontein Shear a highly tectonised granodiorite is sporadically developed along strike for some 4 km (grid reference Q3, Annexure 1, and Plate 6). The rock is characterised by large, randomly oriented hornblende porphyroblasts in a schistose matrix of mainly quartz, biotite and hornblende (Fig. 4.34). The rock has an inequant grain size and very prominent foliation. Clusters of quartz form augen up to 5 mm long and ribbons parallel to the foliation. Large grains show undulose extinction whereas small grains form a granoblastic-polygonal texture.

Saussuritized plagioclase forms small subhedral to anhedral grains. Minor, elongate, microcline grains up to 3 mm long are aligned parallel to the foliation.

Blue-green hornblende occurs as porphyroblasts and elongate prisms interleaved with biotite along the foliation. The porphyroblasts are oval-shaped, up to 10mm along their longest axis and where composed of single grains, are twinned. They form poikiloblasts enclosing small rounded quartz grains, and occur as decussate clusters of euhedral, randomly oriented grains.

Euhedral to subhedral olive-green to brown biotite forms laths 1 mm long, lepidoblastically interleaved with hornblende as folia. Aggregates and small anhedral grains of epidote are associated with biotite and hornblende. Zircon and magnetite are accessory. Sericite outlines the granoblastic-polygonal texture in a network of small grains. Irregularly shaped calcite grains are elongated in the foliation.

The strongly developed foliation is due to:

(i) 1 to 2,5 mm thick bands of large quartz and microcline grains. These bands are up to 15 mm long, their continuity being disrupted by amphibole porphyroblasts. Quartz is inequant, the

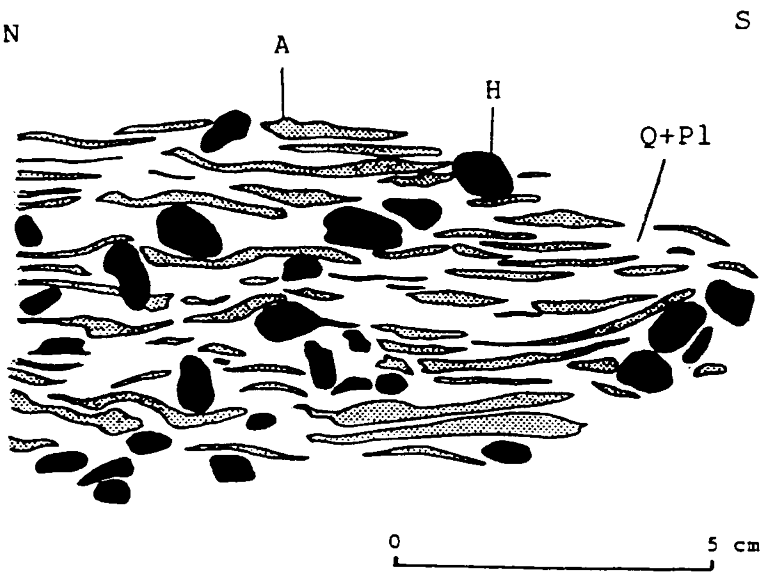


Fig. 4.34 Sheared hornblende gneiss, Groothoek Thrust zone, showing hornblende porphyroblasts (H) in a matrix of quartz and plagioclase (Q+Pl), separated by folia of amphibole and biotite (A). Sense of shear is interpreted as right-lateral, as inferred from asymmetric orientation of amphibole and biotite folia displaced by the hornblende porphyroblasts. The latter overgrow the folia in places. Drawn from a hand specimen.

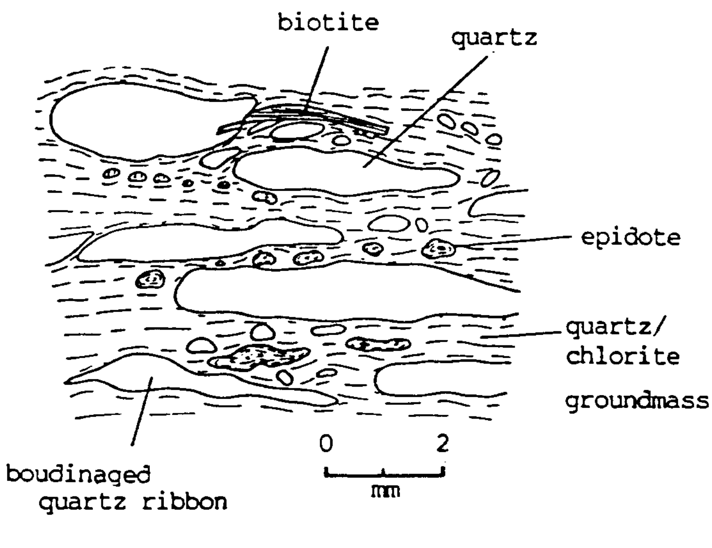


Fig. 4.35 Quartz ribbons separated by a groundmass of quartz, chlorite and epidote, in a mylonitized Vioolsdrif granodiorite. Drawn from a photomicrograph.

large grains measuring 2,5 x 1,5 mm, with undulose extinction. Microcline forms subhedral to anhedral inequant grains;

(ii) bands about 2 mm thick, composed of quartz and plagioclase grains, averaging 0,1 mm in size. These bands have a granoblastic-polygonal texture, individual grains of quartz showing complete recrystallization. Sericite, epidote and sphene are associated with these small grains. Sphene occurs as euhedral to rounded grains, 0,2 mm long and is strongly pleochroic. Calcite forms small irregular grains aligned along the foliation;

(iii) slender interleaving laths of biotite and amphibole together forming folia up to 1 mm thick. These folia characteristically transect the rock at an average spacing of 2 mm apart.

A schistose foliation is developed in the Vioolsdrif granodiorite at contacts with xenoliths of the with Windvlakte suite. A transition from undeformed Vioolsdrif granodiorite to a strongly tectonised rock occurs over a distance of some 100 metres. The development of this tectonised fabric can be traced from zones of lesser deformation, where asymmetric quartz augen progressively change to elongate ribbons in zones close to the contact with the xenoliths (Fig. 4.35). Quartz ribbons are separated by folia of biotite and epidote which become more densely spaced with proximity to the contact. Here small laths of brown biotite cluster together in 1 mm thick folia, and epidote forms elongate grains up to 5 mm long. Mineral assemblages indicate that metamorphic conditions during deformation of the granodiorite were in lower amphibolite facies.

The leucogranite immediately south, and associated with the mafic/ultramafic rocks of the Klipbok complex, has a prominent foliation enhanced by inequant quartz and feldspar grains, and parallel muscovite laths. Mylonites develop at the contact with Klipbok complex mafic rocks, by grain refinement and recrystallization of quartz into ribbons (Plate 16).

Deformation of mafic/ultramafic rocks is not as pronounced as in the siliceous rocks. However, in serpentinites chlorite grains are kinked and have undulose extinction, and sometimes contain associated exsolved ilmenite. Clinozoisite occurs at the top and base of the unit. The eastern-most mafic rock occurrence exhibits a well developed foliation, outlined by lepidoblastically aligned hornblende and epidote laths. The western-most amphibolites are massive, with minor reconstitution of grains. The textural difference is due to the eastern-most occurrence being located closer to a zone of shearing. Amphibole grains in most of the hornblendites display a decussate texture, but in some occurrences hornblende grains have developed parallel to the foliation.

The well developed foliation in the Groothoek Thrust zone is manifested in different ways in different rock-types. Stereographic plots of $S(N)_2$, for example, reveal a consistent west-northwesterly strike with a variable dip to the north-northeast (cf. Subdomain 6.1, Domain 6, Annexure 3). It is also noticeable how the attitude of the S-surfaces within the Groothoek suite to the south of the Groothoek Thrust zone have a divergent orientation compared to those in the zone. This is shown up in structural Subdomain 6.2 where the average strike direction is 320° as opposed to 295° in the Groothoek Thrust zone. This evidence points to a zone of dislocation somewhere between the above mentioned two structural subdomains, i.e., probably at the base of the Groothoek Thrust.

(ii) Lineation

A stretching lineation ($L(N)_2$), shown by the down-dip alignment of mineral aggregates and elongate cordierite and hornblende growth (cf. Plate 4), is well developed along the Groothoek Thrust zone. It has a trend towards 020° and a plunge of 43° in the zone of Groothoek thrusting (see, for example, Subdomain 6.1, Domain 6, Annexure 3).

In the zone immediately south of the Groothoek Thrust $L(N)_2$ trends 034° and plunges 35° to the north-east (see, for example, Subdomain 6.2, Domain 6, Annexure 3). This orientation is slightly oblique to $L(N)_2$ in the Groothoek Thrust zone. The divergence is even more marked in structural Subdomains 2.1 and 2.2.

(iii) Fold orientation

In the Groothoek Thrust zone folds have been attenuated and flattened (Fig. 4.36). Their axial planes have the same orientation as $S(N)_2$, and their fold axes plunge down-dip with a similar north-northeasterly orientation to $L(N)_2$ (see Subdomain 6.1, Domain 6, Annexure 3). Rare flattened folds of this nature occur in the metapsammitic and metapelitic units, while microscale isoclinal folds are present in the easternmost amphibolite (Fig. 4.37).

In the zone immediately south of the Groothoek Thrust zone fold axes of small isoclinal folds have a variable orientation, plotting between azimuths of 330° to 090° (see Subdomain 6.2, Domain 6, Annexure 3). This inconsistent fold axis orientation contrasts with that of the Groothoek Thrust zone.

(iii) mylonites

Mylonites are developed in the leucogranite on contact with rocks of the Klipbok complex (Plate 16). They are also developed in

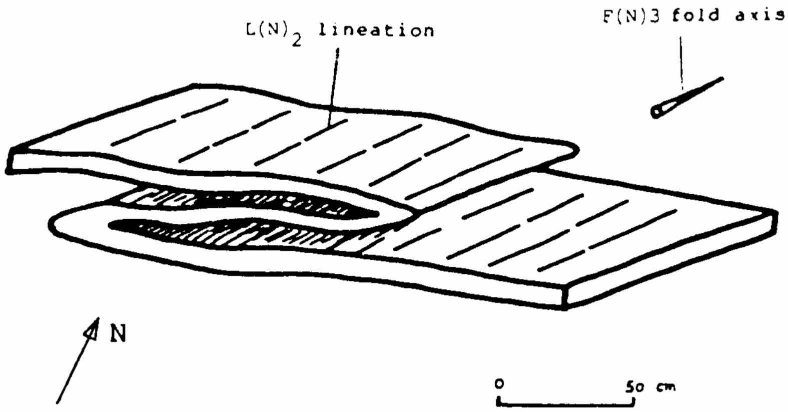


Fig. 4.36 Schematic drawing showing flattened fold, with F(N)3 fold axis and $L(N)_2$ lineation plunging down-dip, Groothoek Thrust zone.

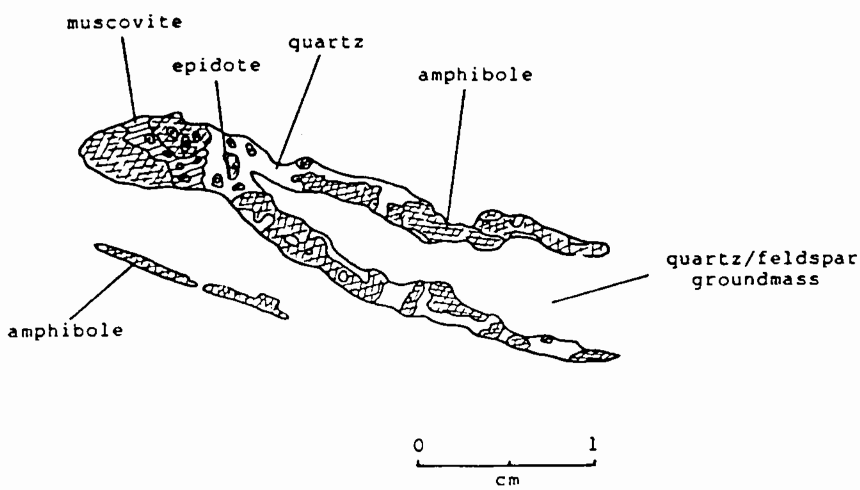


Fig. 4.37 Isoclinal microfold, within mafic unit associated with the Groothoek Thrust zone. Drawn from a thin section.

Vioolsdrif granodiorite on contact with supracrustals and in lithologies of the Groothoek suite.

(iv) Wedge-shaped units in Groothoek suite

A divergence of $S(N)_2$ strike in the Groothoek suite from northwest to west-northwest, just east of Kouefontein Shear, outlines a triangular shape in plan (grid reference O3 and P3, Annexure 1). This anomalous feature may have a tectonic origin (see interpretation to follow).

4.5.2.3 Interpretation of field and petrographic observations

(i) Field data

The discrepancy in orientation of fold axes in the Groothoek Thrust zone and that to its south is interpreted as follows. Fold axes in the Groothoek Thrust zone were rotated in zones of maximum shear strain until near parallel to $L(N)_2$. Down-dip and obliquely plunging fold axes were therefore reorientated in a shear zone, through a simple shear mechanism.

If this interpretation is correct then it presupposes that the majority of folds develop in response to simple shear in the thrusts. In this case folds above the thrust plane have their axes at right angles to the tectonic transport direction whereas folds closer to the thrust plane are rotated into various orientations during thrust movement, depending on the footwall topology (Butler, 1982a; b). Those folds within the thrust zone rotate the most. Their fold axes point down-dip and, depending on the amount of shear, can be oriented parallel to the tectonic transport direction (Bryant and Reed, 1969; Williams, 1977; Cobbold and Quinquis, 1980; Mattauer et al., 1981; Rattey and Sanderson, 1982).

In a zone of simple shear strain fold axes are rotated in the same way as any pre-existing lineation towards the direction of tectonic transport (Sanderson, 1972; Williams, 1978). This mechanism produces fabric elongation parallel to the axial plane, developing ellipsoids with large X/Z strain ratios characteristic of intense shear deformation. The mechanism accounts for down-dip and obliquely plunging fold axes in areas of major thrusting in orogenic belts. Examples include thrust sheets in the Grandfather Mountain Window, southern Appalachians (Bryant and Reed, 1969), the Wood-roffe Thrust, Australia (Bell, 1978), thrusting related to obduction in Corsica (Mattauer et al., 1981) and thrusting in the Variscan of south-west England (Rattey and Sanderson, 1982).

The consistent north-northeasterly trend of $L(N)_2$ within the Groothoek thrust zone suggests that tectonism was concentrated over a broad zone, movement being towards the south-southwest.

The wedge shaped structure in the Groothoek suite might represent a lateral ramp structure, although one continuous thrust plane could not be confirmed in the field. There are, however, minor transecting structures present. This makes it more likely that the wedge-shaped structure can be considered part of an imbricate package.

Wedge-shaped structures in Groothoek suite rocks are probably related to the Ratelfontein Thrust to the south, in view of closer proximity of such structures to it.

(ii) Hornblende porphyroblasts

A hand specimen of deformed Vioolsdrif granodiorite containing prominent hornblende porphyroblasts shows the sense of shear can be interpreted as right-lateral (facing east) (Plate 6). The inferred north to south tectonic transport direction is based largely on the overall asymmetric orientation of hornblende porphyroblasts (Fig. 4.34).

Away from zones of intense shearing hornblende porphyroblasts in Vioolsdrif granodiorite grow as isolated entities within a foliation outlined by biotite (Fig. 4.38). Close to zones of intense shearing decussate clusters of hornblende make large porphyroblasts. Porphyroblast growth in this case appears related to intensity of tectonism, i.e., the larger porphyroblasts develop in zones of greater inferred shear strain, and are randomly oriented (cf. Plate 6, and Fig. 4.39).

The development of this texture is explained by a shear strain mechanism referred to as "deformation partitioning" by Bell et al. (1986). The mechanism incorporates simultaneous shearing and shortening strain, and develops as follows (Fig. 4.40). Heterogeneous deformation taking place in a rock produces undulating, sigmoidal shaped cleavages, due to competency contrasts both within, and surrounding the rock (Fig. 4.40(a)). The elliptical-shaped area, A, is a zone of no strain, whereas surrounding it is a zone of shortening strain, B. Both shearing and flattening take place in zone C.

Quartz and/or, feldspar are deformed into oval-shaped grains in the shortening strain field, whereas phyllosilicates enclose them in the zone of shearing strain, forming an anastomosing cleavage (Fig. 4.40(b)).

In the zone surrounding an ellipsoidal grain, shear strain is aided by slippage along the basal (001) cleavage planes in e.g., the micas and graphite (Fig. 4.40(c)). This is also the zone of dissolution of other minerals, facilitated by shearing. Continued deformation gives rise to microfractures in the elliptical-shaped

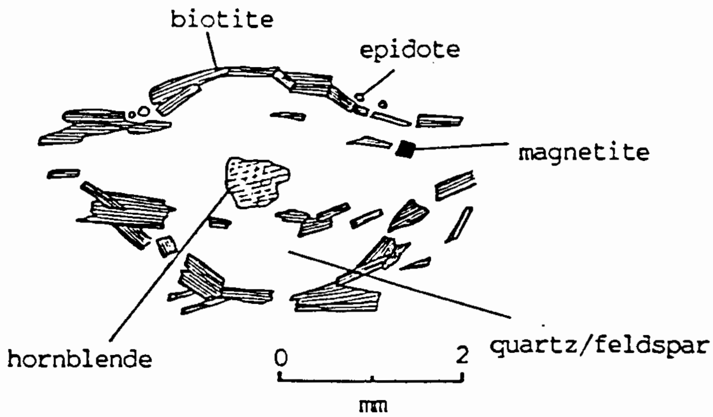


Fig. 4.38 Incipient hornblende porphyroblast develops within a foliation outlined by biotite. Drawn from a thin section of Vioolsdrif Granodiorite, near southern contact.

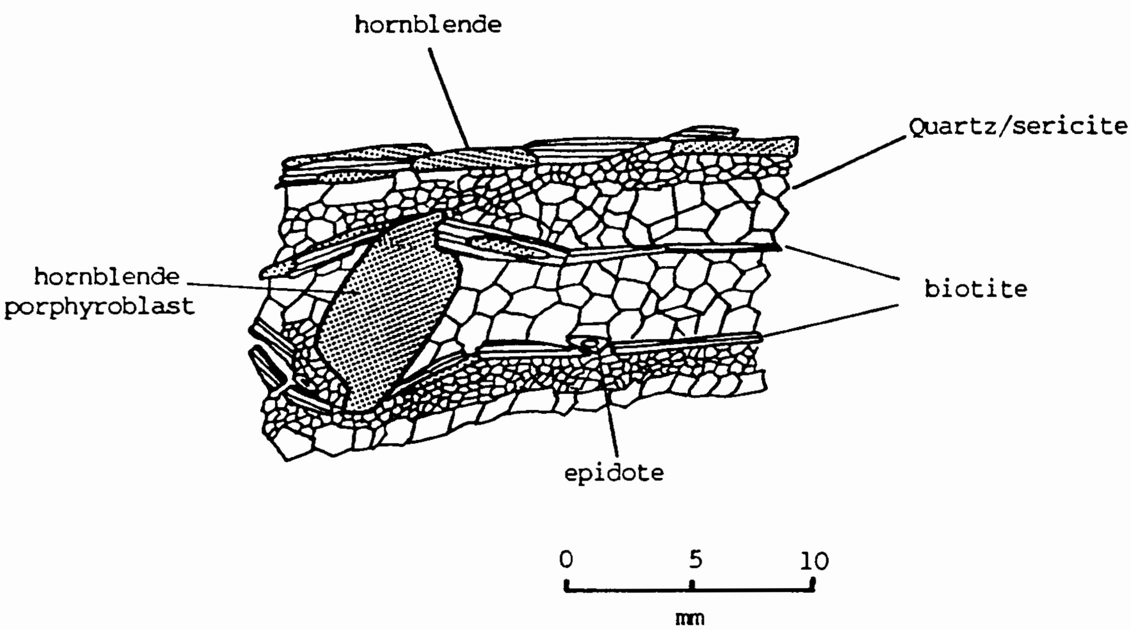
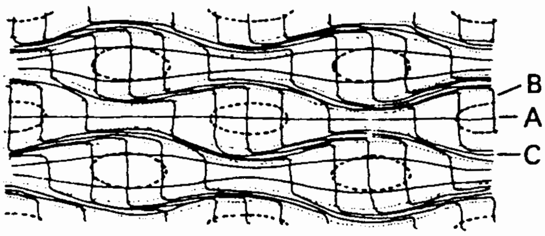
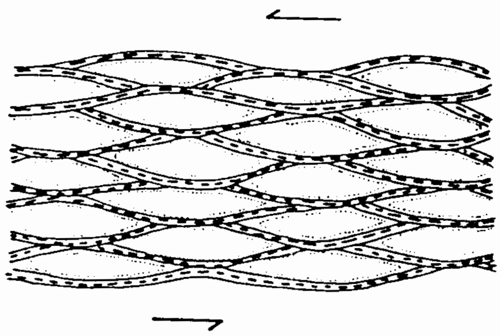


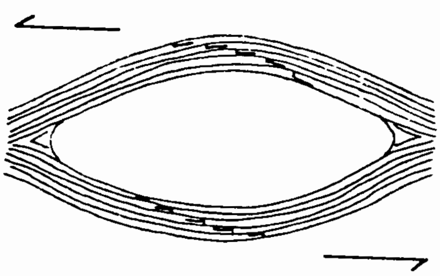
Fig. 4.39 Texture of porphyroblastic hornblende gneiss, southern margin of Vioolsdrif Granodiorite. Note large hornblende porphyroblast oriented at about 50° to foliation. Drawn from a photomicrograph.



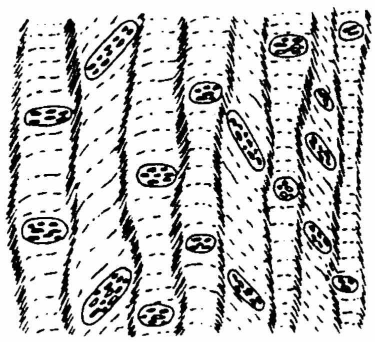
(a) Strain-field diagram showing areas of no strain (A), shortening strain (B) and both shearing and shortening strain (C).



(b) Sketch indicating phyllosilicates wrapping around e.g. quartz and feldspar grains. Phyllosilicates deform by simple shearing whereas quartz and feldspar grains deform predominantly by pure shear.



(c) An ellipsoidal grain surrounded by phyllosilicates. Shear couple shown results in deformation along (001) crystallographic planes of the phyllosilicates.



(d) Development of porphyroblasts within zones of progressive shortening. Longest porphyroblasts grow where earlier foliation is at a low angle to the crenulation cleavage.

Fig. 4.40 Deformation partitioning - a shear mechanism to explain porphyroblast development in sheared rocks (after Bell et al., 1986; a & b above = Fig 1a & b of Bell et al., c = Fig 5c, d = Fig 17). For explanation see text.

grain where porphyroblasts can grow. Chemical potential gradients set up in, for example, shear zones can drive ionic dissolution and solution transfer, thus promoting precipitation and growth of porphyroblasts, particularly in zones of shortening strain.

However, if a younger crenulation cleavage transects the rock, porphyroblasts rotate and lie obliquely to it (Fig. 4.40(d)). Porphyroblasts at an acute angle to the cleavage have an elliptical habit, whereas those at right angles to it are less elongate. Their shape here may be related to the orientation of the older foliation, an elongate habit being favoured where the older foliation is at an acute angle to the crenulation cleavage (Fig. 4.40(d)).

The deformation partitioning theory of Bell et al. (1986) may have some application in explaining the texture in deformed Vioolsdrif granodiorite. However, metamorphism has probably outlasted deformation in this example thus causing hornblende porphyroblast overgrowth across the foliation (Chapter 5).

(iii) Mylonite development

Petrographic evidence suggests that formation of mylonites in the leucogranite is by quartz grain refinement and ribbon development (cf. Plate 16). The preferential occurrence of mylonites in siliceous, as opposed to mafic and ultramafic rocks of the Klipbok complex, is probably due to the competency contrast between them (Waters and Campbell, 1935). These mylonites were probably formed in shear zones as a result of inhomogeneous deformation and strain softening, during conditions of ductile deformation as advocated by White et al. (1980).

Mylonites in the Vioolsdrif granodiorite contact zone west of Kliphoogte exhibit quartz ribbons separated by needle-like folia of muscovite. Plagioclase grains are kinked. These deformation characteristics form under upper greenschist to lower amphibolite facies metamorphism (Simpson, 1985).

Deformation in the zone of Groothoek thrusting was probably effected through simple shear, although this cannot be substantiated due to lack of suitable quantitative strain data. Reorientation of fold axes, shearing phenomena, mylonites and the strong alignment of $L(N)_2$ in the Groothoek Thrust zone do however support a model of simple shearing (Escher and Watterson, 1974; see Fig. 4.41).

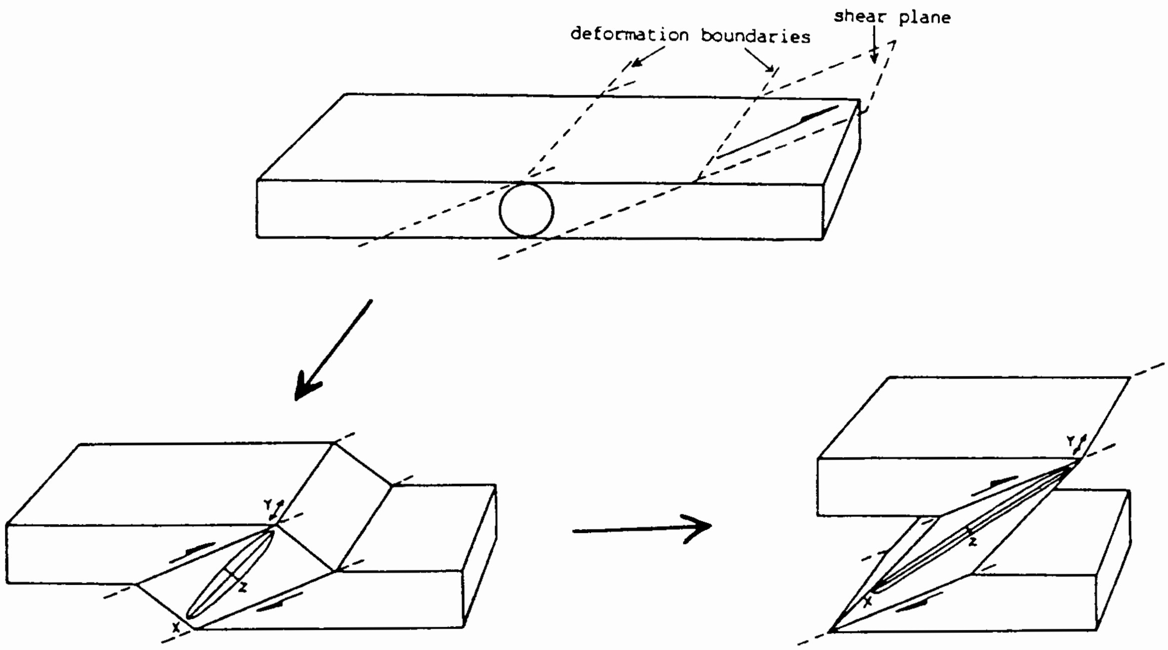


Fig. 4.41 Progressive simple shear deformation in a shear zone (after Escher and Watterson, 1974).

Thrusting appears to have taken place under ductile deformation conditions, favouring a broad zone of deformation (cf. Fig. 4.4). This makes it difficult to locate a definite thrust surface, but rather implies a thrust zone of at least 2 km width. Problems in locating the exact plane of suturing are due to the fact that a variety of high strain zones exist in older, ductile zones of basement shearing (Dewey, 1977). This is the case in the Groothoek Thrust zone where variations in deformation intensity were probably brought about by the diversity of rock-types, and their differing behaviour patterns during tectonism.

An analogy from the western Himalaya, the Main Central Thrust, is not a single plane of thrusting, but a zone of shearing several km thick (Searle and Fryer, 1986). In this case the exact placing of the thrust surface is complicated by metamorphic overprinting (Sinha Roy, 1982). In the Assynt District of Scotland, thrusts similarly do not have a distinct surface, but form a zone of shearing at least 100 m thick (Christie, 1963).

4.5.2.4 Displacement

A 75 km southward displacement along the Groothoek Thrust is calculated by Blignault et al. (1983, p. 22) in the area of the Namaqua geotraverse, based on Ramsay and Graham's (1970) simple shear model of ductile deformation in shear zones. This technique was applied on the hanging-wall foliation in Vioolsdrif granitoids, and the orientation of stretching lineations used as the movement direction. The validity of the method in this case is challenged (C.J.H. Hartnady, personal communication, 1986) because it assumes continuous ductile deformation in a zone where there is evidence in the Groothoek schists showing superimposition of a younger fabric on an older one (Theart, 1980). Application of the Ramsay and Graham (1970) technique in the latter case would not necessarily reflect earlier or perhaps synchronous displacements which may have taken place within, for example, the schistose footwall rocks of the thrust plane.

A modified version of Elliott's (1976) "bow and arrow" rule may be more relevant in the estimation of at least a minimum displacement along the Groothoek Thrust. This requires that

(i) the thrust tip points must be known so that thrust termination along strike can be demarcated (Hossack, 1983), and

(ii) the surface trace of the thrust must be known, so that extremities of displacement at right angles to the strike of the thrust are located in plan.

A maximum displacement is measured, based on a linear relationship between the amount of displacement, and the strike length of the thrust fault (Elliott, 1976). Displacement increases from zero at the tip points until it reaches a maximum along a line at right angles and midway between the two tip points, the maximum amount being measured at the apex of the "bow".

In the area of the Namaqua geotraverse and eastwards the Groothoek Thrust trace is not precisely defined because of paucity of lithological exposure (Blignault et al., 1983; Joubert, 1984; Stowe, 1984; Strydom and Visser, 1986; Strydom, 1987).

Fig. 4.42 shows continuity of the Groothoek Thrust from the study area to the Pofadder lineament. This drawing compiled by the writer is based on the premise that the Wortel Belt mafic and ultramafic bodies outcropping in the Pofadder area have lithological and tectonic continuity with those outcropping in the southern Richtersveld. This premise is justified in view of the linear arrangement of mafic and ultramafic rocks, and their associated tectonism (P. Joubert and C.J.H. Hartnady, personal communication, 1983).

The Groothoek Thrust outcrop trace was delineated from available geological maps spanning the area between Springbok and Pofadder. Mafic and ultramafic outcrops along the Wortel Belt were joined up, and this line extended westwards to link up with the thrusts

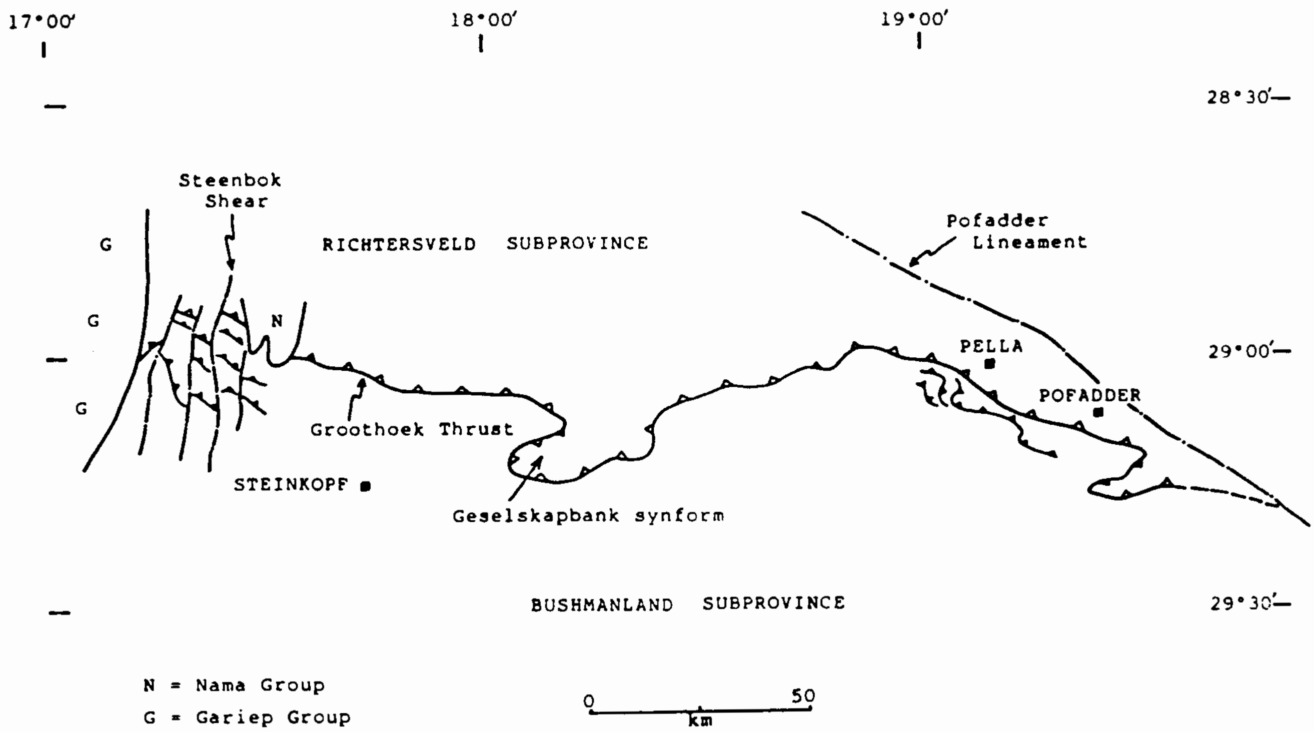


Fig. 4.42 Regional outcrop trace of Groothoek Thrust zone (traced from information provided in regional geological maps; eastern sector modified by M. Watkeys). Dip of thrust variable along its length.

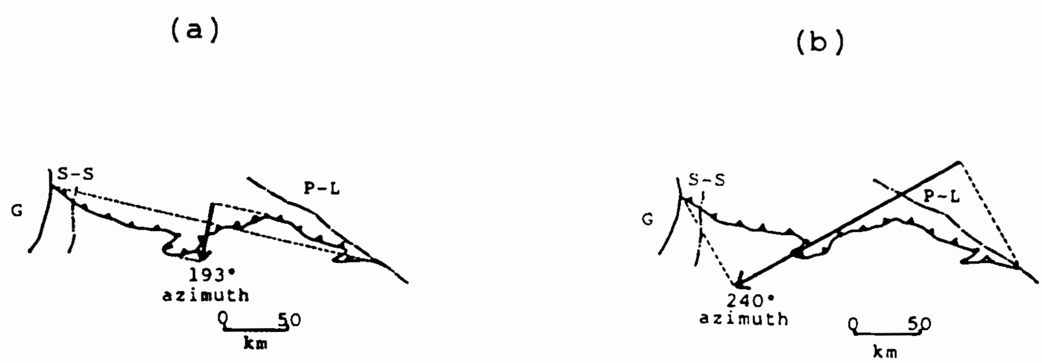


Fig. 4.43 Determination of displacement along Groothoek Thrust, applying modified version of bow and arrow rule (after Elliott, 1976) in two extreme cases:
(a) 50 km displacement on an azimuth of 193°.
(b) 200 km displacement on an azimuth of 240°.

in the Geselskapbank Synform (Strydom, 1985). The position of the Groothoek Thrust here was slightly modified from its position on the 1:100 000 published map (Strydom, 1982) to pass between Vioolsdrif/Haib and Bushmanland lithologies. From the Geselskapbank area the Groothoek Thrust was traced westwards to include the outcrop of mafic bodies in the Safnek area (Ward, 1977; Theart, 1980), and the Klipbok complex (present survey). Portions of the thrust trace in the Pofadder area have been modified from updated information provided by M. Watkeys (personal communication, 1985).

A problem arises as to the placing of the thrust's tip points, due to tectonic and metamorphic overprinting in the critical areas. The western tip point is assumed to be positioned at the contact of basement rocks with the Stinkfontein Formation. This is because overprinting of $S(N)_2$ fabrics by Pan-African $S(P)_2$ makes it difficult to trace the continuity of the thrust much farther west than the Kromnek Shear. The eastern tip point is positioned at the intersection with the Pofadder Lineament, some 50 km south-east of Pofadder. This may not be justified because right-lateral strike-slip movement along the Pofadder lineament has probably displaced the true tip point farther southwards. Establishing the tip point here is therefore difficult, especially as lithologies in this eastern sector are strongly sheared by this post- Groothoek fault (P. Joubert, personal communication, 1985).

Applying the bow and arrow technique in this case nevertheless gives some order of magnitude in the estimation of displacement along the Groothoek Thrust. If stereogram data of $L(N)_2$ stretching lineations from the westernmost area are used (Subdomain 6.1, Domain 6, Annexure 3), then an azimuth of 193° yields a displacement of 50 km (Fig. 4.43(a)). $L(N)_2$ just south of the Groothoek Thrust zone has an azimuth of 240°, yielding a displacement of 200 km (Fig. 4.43(b)). These two conditions represent the two extreme cases of displacement along the thrust. Displacement towards a 190° azimuth is in the writer's opinion more appropriate, in view of field evidence and stereogram data associated with the westernmost extremity of the Groothoek Thrust. The west-southwesterly azimuth (240°) relates more appropriately to the regional foliation and thrusts south of the Groothoek Thrust (section 4.5.3).

The 50 km displacement estimate is probably a minimum value; this distance being measured at the apex of the "bow", i.e. in the Geselskapbank area some 100 km east of the study area. Accurate measurement of displacement using this technique is not possible because of difficulties in defining the tip points of the thrust. If the figure of 50 km maximum displacement along the Groothoek Thrust is accepted as an approximation, then this does mean that displacement in the study area is in fact somewhat less than this figure because the area lies close to the western tip point. However, no account has been taken of modifications of the Groothoek Thrust surface by folding in the Geselskapbank synform, nor of bulk shortening, for example. These cautionary notes further show that the above-mentioned 50 km figure for displacement along the Groothoek Thrust is to be accepted as an approximation.

4.5.3 Ratelfontein Thrust

This thrust zone occurs at the base of the Ratelfontein suite, and is continuous across the area (cf. Fig. 4.31). Thrust planes are present in schistose rocks overlying the basal quartzites, and higher in the sequence (Plate 17). Changes in dip angle across strike in schists of the overlying Ratelfontein and Groothoek suites characterize the zone. These features are interpreted as part of an imbricate thrust package; the amount of dip generally increases towards the north.

Thrusting is localized in metapelites and meta-quartzites (cf. Plate 2). The metapelite here has an irregular shape and thickness, as shown in the eastern sector of the photograph. I interpret this irregular configuration as the product of thrusting; most of the movement being taken up in the metapelite unit (Fig. 4.44). Thrust movement directions are depicted on the diagram relative to a projected thrust ramp below the metapelite. This suggests that movement of the hanging-wall was towards the south. This configuration strongly resembles Gwinn's (1964) sketch showing the geometry of folded strata and associated faulting in the Central Appalachian region (Fig. 4.45). Gwinn attributes these structures to concealed décollement zones in the "thick skinned" central Appalachians. A significant effect of this deformation is that the strata are thickened in section (Figs. 4.44 and 4.45).

Most of the Ratelfontein suite quartzites have boudinaged shapes, and are arranged en echelon in section (cf. Fig. 3.7) but occasionally a tectonic banding is present. On a mesoscale the orientation of the banding in the basal quartzites is consistent with that of imbricate thrusts in the muscovite schists overlying the quartzites. I attribute the segmented character and boudinaged quartzite shapes to two possible mechanisms:
(i) Segmentation of a competent bed into boudins during simple shear deformation. En echelon boudin structures can develop in this way (Ramsay and Huber, 1983, Fig. 2.14B; Cobbold and Quinquis, 1980; Needham, 1987);
(ii) Modification of the boudins into "pip" shapes through re-orientation of the stress field during further deformation. Boudinaging in the Trout Lake area, Saskatchewan, is attributed to a similar cause (Schwerdner, 1970, p. 894).

The position of the main décollement zone is below the basal quartzite unit, i.e. at the base of the Ratelfontein suite. Of significance here is that the Kouefontein granite gneiss occupies this stratigraphic position, indicating that it has exploited this

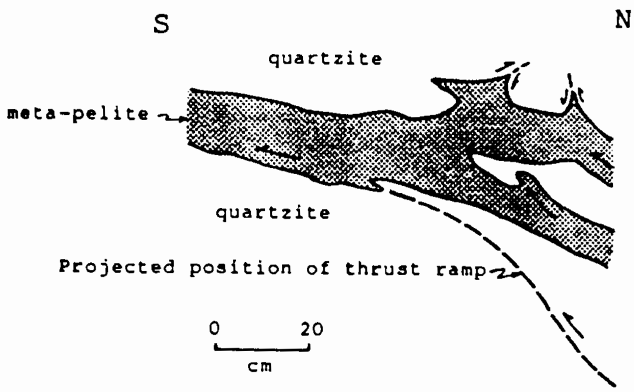


Fig. 4.44 Interpretation of thrust movement in metapelite bed, Ratelfontein suite. Diagram traced from Plate 2.

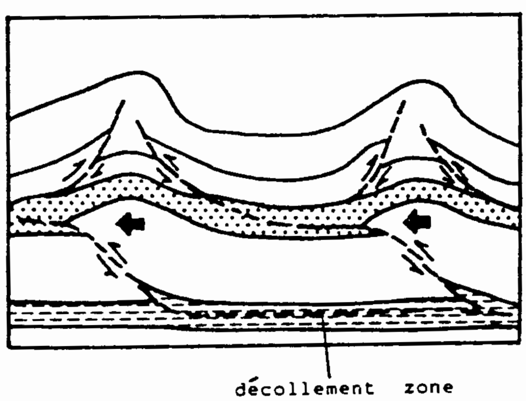


Fig. 4.45 Schematic illustration from the Central Appalachians showing the development of reverse faults in relation to folds and décollement zones (after Gwinn, 1964, Fig. 19). Note thickening of the section above thrust ramp position.

structural plane across the entire area.

Defining the Ratelfontein Thrust west of Steenbok Shear is speculative for two reasons. Firstly, metamorphism associated with the Pan-African event has overprinted and obscured evidence of thrusting, although a well developed schistosity is still evident in metapelitic rocks near the Stinkfontein Formation unconformity (cf. Fig. 3.10). Secondly, continuity of the Ratelfontein Thrust west of Kromnek Shear in an area of poor outcrop is based on the interpretation that the quartzite boudins define a major zone of dislocation across the entire area.

The relative timing of deformation along this thrust zone is speculative. However, based on the geological history, I propose that the Ratelfontein Thrust represents a zone of simple shear deformation. Thrusting caused imbrication in schists of the Ratelfontein suite, minor thrusts in the quartzites, and en echelon rearrangement of quartzite boudins. This was followed by intrusion of the Kouefontein granite gneiss.

Tectonic movement was towards 214°, as observed from orientation of $S(N)_2$ and $L(N)_2$ in the Ratelfontein Thrust sheet. It occurred at a slightly shallower angle of dip compared to the Groothoek Thrust zone (Subdomain 6.2, Domain 6, Annexure 3).

4.5.4 Kouefontein North Ramp Structure

This structure is confined to the narrow crustal block between the Steenbok and Kouefontein Shears (Fig. 4.31, thrust zone 4). It has a north-westerly strike (340°) and a north-easterly 60° dip. The ramp is a strongly tectonised 1 m wide zone of quartz-sericite schist. A "mullion"-type lineation plunges steeply towards the north-east in the shear zone.

Just above the zone of shearing folds with amplitudes of about 50 cm are present. Their fold axes trend parallel to the zone of shearing, plunging shallowly to the north-northwest. The folds display a variable range of interlimb angles (cf. Park, 1983, p. 13). Tight folds are closest to the zone of shearing, but these give way to open and ultimately monoclinal folds away from the zone of shearing (see Fig. 4.46).

Mylonitic textures in quartz-sericite schist display quartz ribbon structures and muscovite porphyroblasts set in a fine-grained matrix of sericite and quartz (Fig. 4.47). The quartz ribbons have variable widths up to 2 mm, and are segmented into asymmetric augen shapes. Quartz within the ribbon structures mostly shows extreme undulose extinction, with recrystallization of smaller (0,1 mm) grains transecting the ribbon structure at angles of about 30°. An entire ribbon or augen is sometimes composed of recrystal-

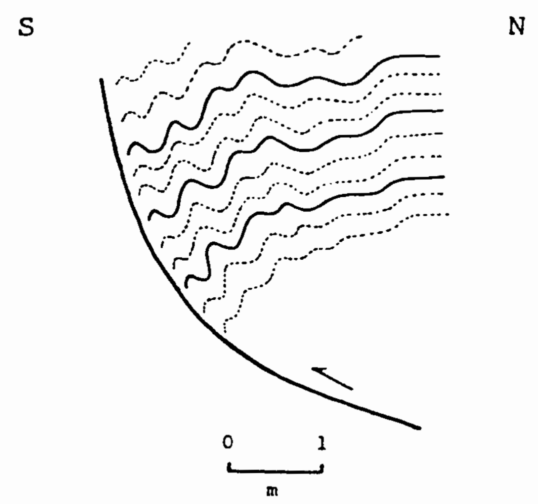


Fig. 4.46 Folds developed in relation to the Kouefontein North Ramp structure. Sketched from field photographs.

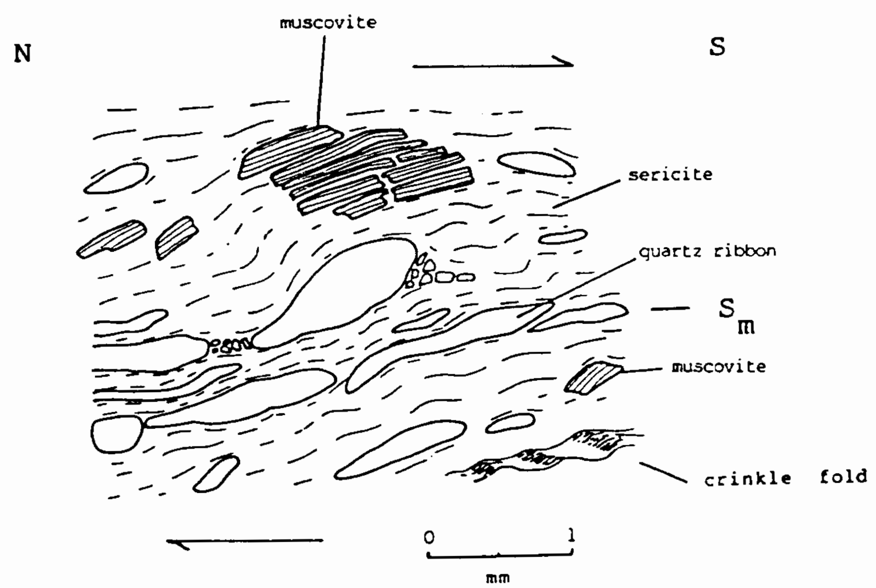


Fig. 4.47 Mylonitic textures in quartz sericite schist, Kouefontein North Ramp structure. Note shape of quartz ribbons/augen, and "fish" muscovite indicate a dextral sense of shear. Drawn from a photomicrograph. Interpretation of movement direction after Simpson and Schmid (1983), and Eisbacher (1970).

lized grains. Muscovite porphyroblasts are deformed into "fish" shapes (Plate 18).

There are three possible interpretations regarding the origin of this ramp structure.
(i) The ramp is an entity formed in response to a north-south oriented stress direction, as its strike is parallel to other Mid-Proterozoic thrust zones (cf. Fig. 4.31).
(ii) It could have developed during the Pan-African tectonic event, in which case its origin as a zone of thrusting is intimately related to steeply-dipping, north-easterly trending shear zones.
(iii) It could be an extensional listric fault, formed in response to crustal thickening during Namaqua thrusting.

An argument against (i) is that the ramp is a very localized structure, traceable over a strike length of less than 1 km. In support of (ii) the ramp may have formed during rotation of the Kouefontein Shear (cf. section 4.3.2.4). This would result in uplift of the hangingwall block, with concomitant development of the ramp and associated fold structures as interpreted in Fig. 4.46. Field mapping reveals that a north-trending shear zone, just east of Steenbok Shear, offsets the Ratelfontein suite by uplifting the eastern block (Annexure 1, grid reference N3). From this position the shear veers towards a south-southeasterly direction, where it appears to splay into the Kouefontein North Ramp structure. The relationship of thrusts to shear zones is well known (de Sitter, 1956), and in this case could explain the origin of the Kouefontein North Ramp structure. However it is possible that interpretation (iii) above is valid because movement directions can also be interpreted as opposite to those depicted in Fig. 4.46 and 4.47 (Passchier, 1986). In this case the ramp structure developed during the Namaqua event. I prefer interpretation (ii) above, based on field evidence, although development of the Kouefontein North structure may well be a Namaqua-related structure. Its origin remains debatable.

4.5.5 Kouefontein mylonite gneiss

The Kouefontein granite gneiss is strongly tectonised towards its lower contact with the Sabieboomrante adamellite gneiss, and reveals protomylonitic textures on a microscale. This relatively broad zone of sheared rocks is located just east of Kouefontein Shear (cf. Fig. 4.31, thrust zone 5). It is also sporadically developed in places west of Steenbok Shear, although this is not shown on the diagram.

Most of the quartz grains have undergone plastic deformation, while some have survived as large inequant grains with pronounced undulose extinction, and sutured boundaries between them. Small polygonal, recrystallized quartz grains surround large quartz and microcline grains in a typical mortar texture. These define narrow asymmetric zones of strain which surround the larger grains in a protomylonitic texture (see Fig. 4.48).

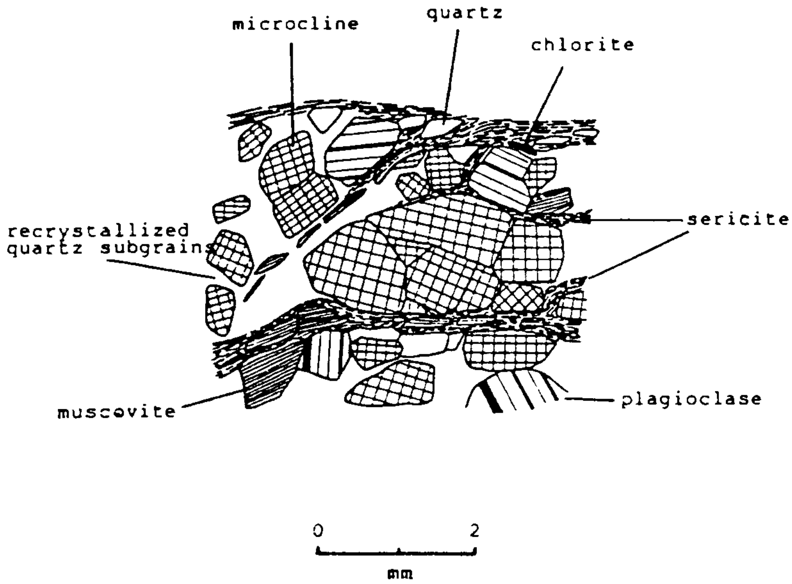


Fig. 4.48 Mylonite gneiss texture in Kouefontein granite gneiss. Note how microcline augen are outlined by sericite and recrystallized quartz grains. Sketched from a thin section.

Large microcline grains show embayed boundaries along which re-crystallized feldspar and quartz grains are present. Brittle deformation textures such as microfracturing in plagioclase are also present. Sericite outlines the asymmetric augen shape of the large microcline grains.

These protomylonitic textures in the Kouefontein granite gneiss are interpreted as forming under upper greenschist, or perhaps lower amphibolite facies of metamorphism.

Fabric data in the Kouefontein granite gneiss reveal that $S(N)_2$ has a strike orientation slightly west of that to the immediate north (Subdomain 6.3, Domain 6, Annexure 3). $L(N)_2$ plots as two maxima, on bearings of 030° and 060°. The interpretation here is that the Groothoek Thrust trend has been superimposed on an older 060° trend which is more prevalent in the southern portion of the study area.

4.5.6 Sabieboomrante Thrust

This sporadically developed thrust occurs near, or at the top of the Sabieboomrante adamellite gneiss. It is continuous across the northern contact of the Skurwehoogte granitoid to the west, and terminates just west of Tierkloof Shear (cf. Fig. 4.31).

A tectonised fabric with a well developed foliation and lineation characterizes this zone (Plate 19). The lineation here has a trend and plunge of 045°/53° (Subdomain 6.4, Domain 6, Annexure 3; and Fig. 4.3), compared to that of the Groothoek Thrust zone which is north-north-easterly oriented.

Textures reveal inequant shapes of feldspar and quartz grains, the latter showing undulose extinction and ribbon development. Augen-shaped feldspar and quartz grains are sometimes imbricated, the individual grains separated by sutured boundaries. These are located between a groundmass of small, recrystallized bands of quartz with associated plagioclase, epidote, muscovite and apatite. Small quartz grains show high-angle boundaries against large grains, indicating some recrystallization and recovery. Olive-green biotite forms broad laths parallel to the ribbon foliation and, together with narrow muscovite laths outlines a cleavage which encloses large feldspar and quartz augen. Chlorite replaces biotite along grain edges. Anhedral sphene and epidote grow across most of the other grains.

Myrmekite is significant in the sheared rocks in that it is strain related. It forms in embayments along large K-feldspar (microcline) grains (Fig. 4.49). Vermicular intergrowths are better developed along the edges of the feldspar grains facing the shortening direction.

I interpret the sense of shear as right-lateral (facing east) from the orientation of asymmetric feldspar augen. This infers a north over south movement direction along the Sabieboomrante Thrust. Myrmekite here is interpreted as forming under amphibolite facies metamorphism, similar to that occurring in granitic rocks from the Borrego Springs-Santa Rosa mylonite zone in southern California (Simpson, 1985). In the latter example it occurs along the edges of augen shaped orthoclase grains and has grown during an S-C cleavage development, being more abundant along the sides of orthoclase grains affected by the most strain (Fig. 4.50).

Takagi (1986) describes similar myrmekitic textures from mylonites in the Median Tectonic Line, Central Japan. Myrmekite here is intimately related to mylonite development, where K-feldspar augen control the fluxion structure, while fine-grained myrmekite aggregates form along the augen edges.

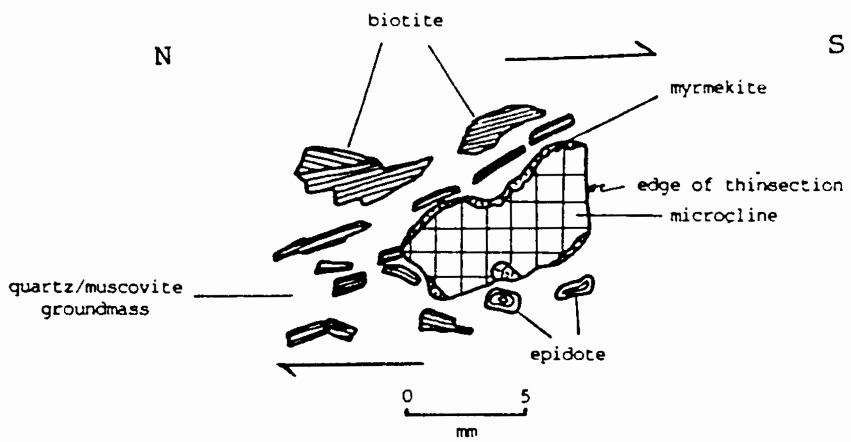


Fig. 4.49 Myrmekite development along edge of microcline grain, within Sabieboomrante Thrust. A dextral sense of shear is interpreted here from the greater abundance of myrmekite on the edge of microcline augen facing the shortening direction. Drawn from a photomicrograph.

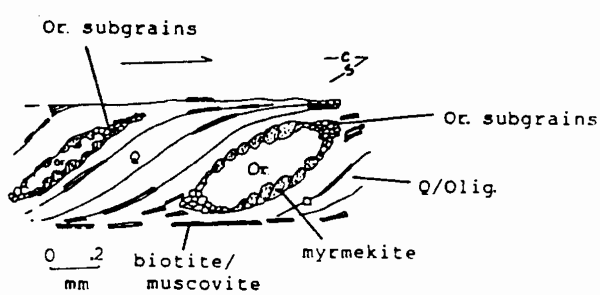


Fig. 4.50 Strain-related myrmekite along edges of orthoclase grains. This texture develops in relation to S-C cleavages, in granitic rocks from the Borrego Springs - Santa Rosa mylonite zone, Southern California (after Simpson, 1985, Fig. 4).

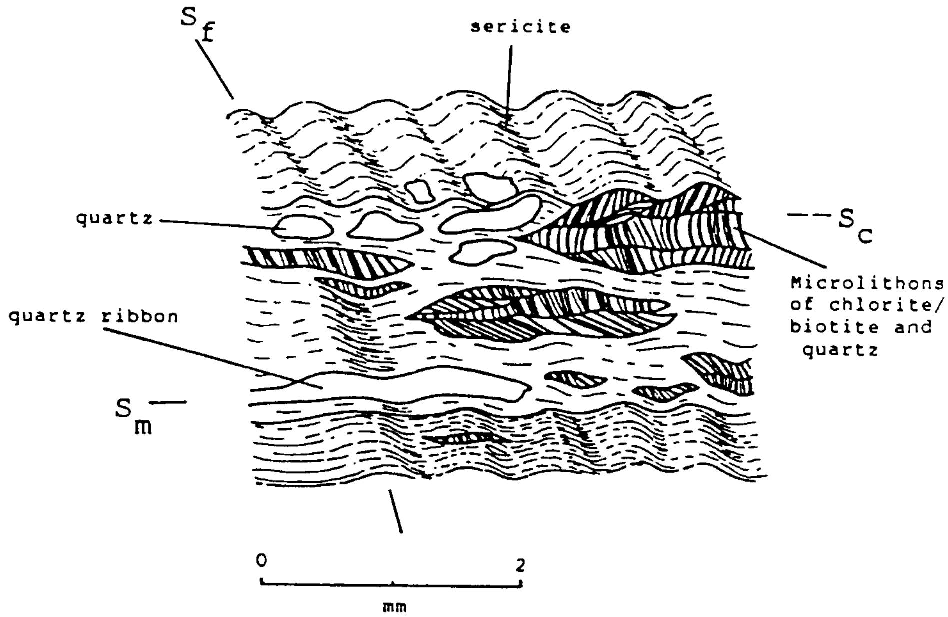
4.5.7 Chabiesies South Thrust

This thrust zone occurs in the south-eastern sector of the study area, where the Chabiesies suite has a divergent strike compared to the Ratelfontein suite (Fig. 4.31, thrust 7). Just north of the thrust zone $S(N)_2$ has a strike of 287°, whereas to the south of it the average strike is 330° (Subdomains 8.2 and 8.3, Domain 8, Annexure 3). $L(N)_2$ trends in the latter area are oriented clock-wise to those farther north (Fig. 4.3). The thrust crops out over a 20 metre wide quartz chlorite schist zone dipping steeply to the north-northeast, just over 1 km south-west of the old Chabiesies homestead (grid reference Q11, Annexure 1). Associated with this outcrop are boudinaged quartz rods plunging down-dip, as well as small crinkle folds in the schistose rock. The orientation of fold axes is approximately horizontal.

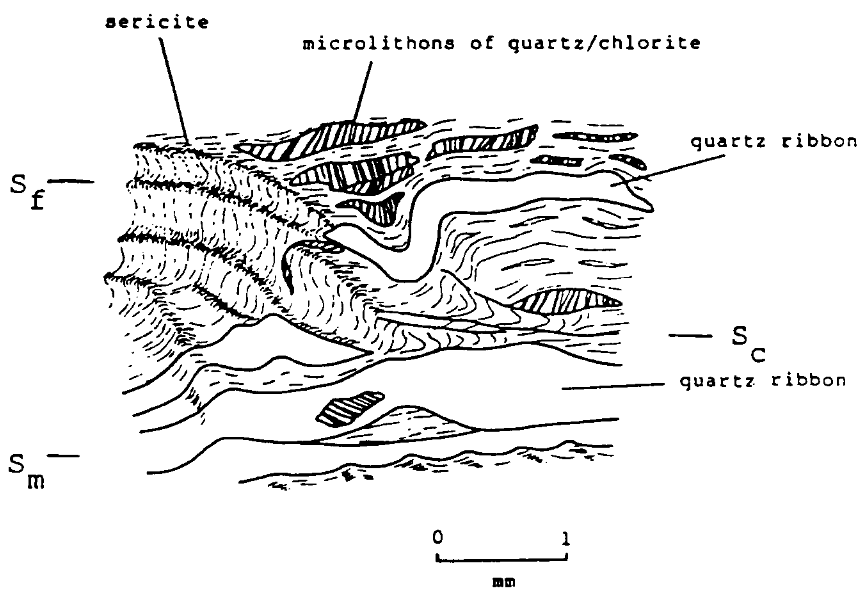
The thrust zone shows retrograde metamorphic effects and two cleavages. Quartz ribbons defining the mylonitic foliation (Sm) are separated by a fine-grained groundmass of sericite, chlorite and quartz (Plate 20, and Fig. 4.51). A younger cleavage (Sc) which is parallel to the quartz ribbons forms microlithons averaging 0,5 mm thick. Skeletal intergrowths of chlorite and quartz, and biotite and quartz are deformed into asymmetric shapes by the latter cleavages.

Many quartz ribbons are stretched into boudin shapes by a later crenulation cleavage (Sf) which has developed at about 50° to the mylonitic foliation. The crenulation cleavage is mostly manifested in microfolds with a wavelength of slightly more than 1 mm. These microfolds are best developed in broad sericitic bands, but they also fold quartz ribbons (Fig. 4.51(a)). In some thin sections the crenulation cleavage (Sf) deforms microfolds into a recumbent orientation with shearing taking place along the upper limb, near parallel to the mylonitic foliation (Fig. 4.51(b)). Here quartz ribbons and microlithon cleavage (Sc) have become tightly folded. From oriented thin sections the crenulation cleavages (Sf) indicate a movement direction from north to south.

Farther east and south-east from this outcrop the thrust zone is obscured by alluvium for some 2.5 km, but reappears as narrow zones in metapsammitic/ pelitic units outcropping in the extreme southeast portion of the area. The metapsammitic/pelitic units are composed mainly of sericite, with some chlorite, chloritoid and remnant garnet, biotite, muscovite, quartz and microcline. Sericite makes up about 60% of the rock, occurring in large clusters where it has formed after feldspar, and as a fibrous matt after sillimanite. Sericite bands are folded into tight and isoclinal microfolds (F(N)4 generation; Fig. 4.52), and correspond to the crenulation cleavage deformation phase in the thrust zone outcrop farther west (cf. Fig. 4.51(b)). Irregularly shaped chloritoid



(a) Three phases of deformation are shown up by quartz ribbons (Sm), microlithons composed of skeletal intergrowths of chlorite/quartz and biotite/quartz (Sc), and a crenulation cleavage (Sf).



(b) Extreme development of a crenulation cleavage (Sf) which assumes a recumbent attitude, and folds the quartz ribbons (Sm), and microlithons (Sc).

Fig. 4.51 Mylonitic textures in chloritic quartz schist, Chabiesies South Thrust. Drawn from photomicrographs.

laths averaging 1 mm long have grown across the fibrous mass of folded sericite, and have subsequently been slightly rotated (Fig. 4.52).

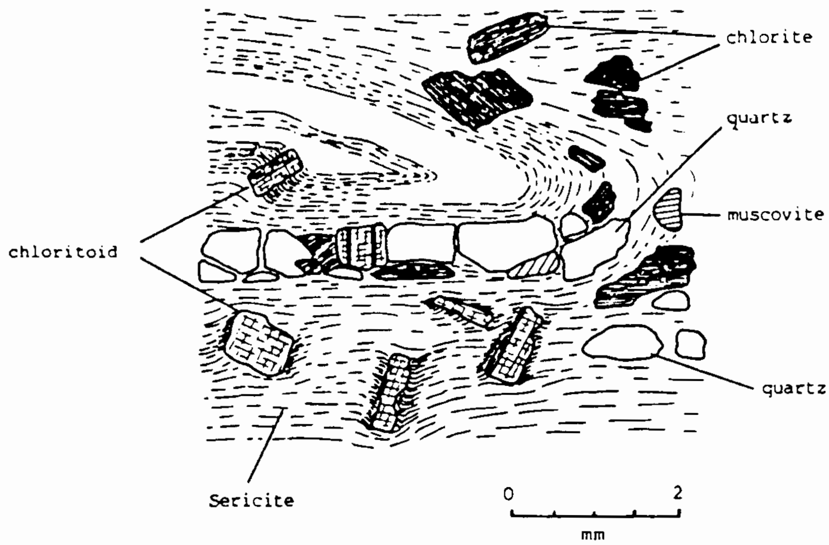


Fig. 4.52 Isoclinal microfold in sericite quartz chlorite schist, Chabiesies South Thrust zone. Note chloritoid laths, slightly rotated. Drawn from photomicrographs.

Continuity of this thrust zone west of Chabiesies is speculative. It may coincide with the southern contact of the Sabieboomrante adamellite gneiss, although it is drawn in south of this position (cf. Fig. 4.31).

4.5.8 Kabies se Berg Thrust

This thrust zone is located in the crustal block between the Kromnek and Tierkloof Shear (Fig. 4.31, thrust zone 8). It separates the Chabiesies suite in the south from the Kouefontein granite gneiss in the north. South of the thrust the lithological strike of the Chabiesies suite is north to north-northeast, although this is modified by open folding (Subdomain 2.4, Domain 2, Annexure 3). North of the thrust the foliation in the Kouefontein granite gneiss strikes east-west, and farther northwards it strikes 310° (Subdomain 2.3, Domain 2, Annexure 3). The position of this thrust zone is therefore interpreted from the disconformable relationship between lithologies of the Chabiesies suite and those to the north. It could be projected eastward to join up with the

Chabiesies South Thrust, but this suggestion remains speculative. The possibility therefore remains that the Kabies se Berg Thrust may represent an older thrust, although branch line relationships which could confirm this relationship are obscured in this area.

4.6 AN OVERVIEW OF SEQUENCE OF EVENTS

The term "event" is here used as being synonymous with a deformation phase which produces a certain set of structures. Several "episodes" of deformation can take place within an event. In the study area at least three events are recognized, viz; Early Proterozoic (D(R)), Mid-Proterozoic Namaqua (D(N)) and the Late Proterozoic/Early Palaeozoic Pan-African event, designated as D(P). Deformation episodes are numbered sequentially (cf. Table 4.1). In a summary of the Early to Mid-Proterozoic history of the north western part of the Namaqua Province a number of other events between the D(R) and D(N) events are proposed, but these will be elaborated on only in Chapter 6.

To meaningfully work out a sequence of events in an orogenic belt structures are correlated relative to each other, and appropriately slotted in with the chronological order of igneous and metamorphic events (Hobbs, Means and Williams, 1976, p. 354; Williams, 1985). In old orogenic belts, e.g. the Namaqua mobile belt, unraveling the sequence of events is made more complex by two factors. Firstly, a span of at least 1000 Ma in geologic time is involved, during which several tectonic events and episodes have occurred. Secondly, igneous events and metamorphic overprinting during the Early to Mid-Proterozoic have obliterated evidence of older tectono-thermal events. This ensures that only the younger events have left their imprint on the rocks, as is the case in other mountain belts, e.g. the Alps and Himalaya. Radiometric dating by Nd and Sm isotope methods is a useful tool for solving problems involving age relationships in complex metamorphic terranes such as occur in Namaqualand (Betton, 1984), but dating by these methods has not been attempted in the study area.

The sequence of deformational events proposed here is based mainly on field evidence. A reliance is placed on radiometric age dates of rock-types common to the present area and other parts of the Namaqua Province (Nicolaysen and Burger, 1965; De Villiers and Burger, 1967; Allsopp et al., 1979; Welke et al., 1979; Reid, 1979a; b; 1981; SACS, 1980; Reid and Barton, 1983).

In the relative dating of Mid-Proterozoic fold structures overprinting relationships similar to those described by Tobisch (1966) and Hobbs, Means and Williams (1976) are used in this study. Generally, Late Proterozoic Pan-African deformational events can be bracketed between the onset of the Gariepian Cycle (Kröner, 1974; 1980), and deformation of the Cambrian Nama Group

(Ritter, 1980). However, in the study area Pan-African tectonism is shown to continue after deposition of the Nama Group and is interpreted as one event - the D(P) event.

4.6.1 Early to Mid-Proterozoic deformation

4.6.1.1 The D(R) event (Eburnian orogeny)

In the study area there are no pre-1900 Ma isoclinal folds in lavas of the Orange River Group similar to those occurring north of the Orange River (Blignault, 1977). However, thrusts, e.g. the Windvlakte Thrust, can probably be assigned to an event related to that proposed above by Blignault (1977), because these structures occur only in the metavolcanic rocks and not in Vioolsdrif granodiorites (1930 Ma) which intrude them.

It is also possible that, either during this episode or soon afterwards, mafic and ultra-mafic rocks of the Klipbok complex were emplaced along a lineament which now separates rocks of the Vioolsdrif Suite from gneissic and schistose rocks of the Groenrivier suite. This aspect will be discussed in Chapter 6.

4.6.1.2 Post D(R) - pre-D(N) events (1800 - 1200 Ma)

In the study area structures developed prior to the Groothoek thrusting episode (Namaqua event at 1200 - 1100 Ma) are largely obliterated. This is because of intense tectonic and metamorphic overprinting during this episode.

The Ratelfontein Thrust is interpreted as the oldest and certainly one of the most important zones of dislocation which developed prior to the Namaqua event. It is transgressive to the lithological strike of the Chabiesies suite, but is separated from the latter by thick granitic sheets (Chapter 3). The main décollement plane is at the base of the Ratelfontein suite just below the boudinaged quartzites. The thrust surface probably merges with the Kabies se Berg Thrust west of Tierkloof Shear, but this is difficult to prove as transecting granitic intrusions obscure possible branch line relationships.

During emplacement of the Ratelfontein Thrust imbricate thrusting took place within the overlying thrust sheet, deforming lithologies of the Ratelfontein, Groothoek and Groenrivier suites and producing a crustal stacking wedge (Plate 17). I propose that the Ratelfontein Thrust sheet thickened during the initial stages of deformation, forming an imbricate stack extending up to the present position of the Groothoek Thrust (Annexure 2).

A phase of magmatic intrusion ensued, the Sabieboomrante adamellite gneiss sheet intruding along the base of the Ratelfontein

Thrust sheet. The southern contact of this thick sheet-like body transgresses the Chabiesies suite obliquely, the latter containing strata of the Gladkop Suite (Annexure 1). This suggests that the sheet was emplaced later than 1800 Ma (cf. van Aswegen, 1988). The Sabieboomrante Thrust near the top of this granitoid body developed simultaneously or soon after intrusion of the pluton.

The Kouefontein granite subsequently intruded as a thick sheet along the same zone, but accessed predominantly along the top of the Sabieboomrante adamellite gneiss. Field evidence shows that a well developed augen texture is present at the base where it is in contact with the Sabieboomrante adamellite gneiss. This texture suggests that the pluton is at least in part syntectonic.

The Kouefontein sheet-like granitic body, and similarly the Sabieboomrante adamellite gneiss have exploited a linear zone of crustal weakness in a manner similar to that described by Pitcher (1978) for granitic bodies in the South American Andes. Pitcher notes that early, large, lensoid-shaped syntectonic granites intrude along major lineaments such as faults in the Andean Mountain range, whereas younger plutons are smaller, and tend rather to be circular in shape. Early Andean plutonism is fault controlled and long lived, and is related to plate action (Pitcher, 1978, p.644). Some analogies regarding intrusion and granite emplacement can therefore be made between granitic bodies of the southern Richtersveld and the Andean mountain chain, although there is a large difference in the ages of these two mountain belts.

4.6.2 Mid-Proterozoic deformation (D(N) (Namaqua) event between 1200 - 1100 Ma)

4.6.2.1 The D(N)1 episode of deformation

The first episode of deformation took place some time prior to the Groothoek thrusting episode and the onset of Namaqua metamorphism. Evidence for this is, however, very limited, for example, it can be found in mesostructures such as $S(N)_1$ in F(N)2 folds, and $S(N)_1$ microfabric in metapelites (cf. Table 4.1). Regional metamorphism during the Namaqua event has completely obscured structural relationships during this episode.

4.6.2.2 The D(N)2 episode of deformation

Compressional tectonism at about 1200 Ma developed F(N)2 folds and the regional foliation $S(N)_2$ during this episode (Table 4.1). Tectonic movement was towards the south-west as shown by the orientation of the pronounced $L(N)_2$ lineation.

4.6.2.3 The D(N)3 episode of deformation

The Groothoek Thrust broke back into the hangingwall north of the Ratelfontein Thrust in response to tectonic movement towards the south-south-west. It formed a predominantly sheared fabric ($S(N)_3$) in the ductile zone of thrusting, and also folded the $S(N)_2$ fabric (section 4.1; Table 4.1). Within the thrust zone the most prominent deformation effects include a well developed west-northwesterly striking foliation, and re-orientation of linear and fold structures which consistently show a south-southwesterly tectonic transport direction. Features of ductile deformation which took place under amphibolite facies conditions along the main tectonic contact zone include mylonite textures and growth of hornblende in Vioolsdrif granodiorite.

The fabric of mafic and ultramafic rocks of the Klipbok complex was reconstituted during this tectonic episode. The latter were subsequently intruded by a leucogranite localized along the thrust zone. This granite could have originated mainly through shear heating during a thrusting episode, similar to the origin of leucogranites associated with major thrusts in the Himalayan and Hercynian orogenic belts (Harris et al., 1986). However, this suggestion is speculative and would need confirmation through detailed geochemical studies.

From orientation of $L(N)_2$ and textural evidence in hornblende gneiss units in the thrust zone (cf. Plate 6) I infer that the volcano/plutonic rocks of the Richtersveld Subprovince were displaced southwards over those of the Groenrivier suite during this tectonic episode. The Groothoek thrusting episode affected lithologies in the foreland for a distance of some 6 km south of the main thrust zone by developing a foliation in the Kouefontein granite gneiss, a stretching lineation (Groothoek trend), and a mylonite gneiss near the base of the Kouefontein pluton.

In attempting to assign an age to the Groothoek thrusting episode constraints are best provided by relative dating of granitic plutons, especially those intrusive into the thrust zone, and those affected by the thrusting. Blignault et al. (1983, p. 22) apply this method in the area of the Namaqua geotraverse near Steinkopf where they assign a syn-Spektakel Suite age (1100 Ma) to the thrusting. This age is based on the observation that the Groothoek Thrust zone deforms the Konkyp augen gneiss, thus making the age of thrusting younger than 1800 Ma (cf. section 2.3.3.1). In the latter area the thrust zone is associated with a zone of retrograde metamorphism and only marginally deforms the Wyepoort granite, a pluton intrusive into the hangingwall of the thrust. The latter is regarded as a correlative of the 1100 Ma Spektakel Suite and therefore an 1100 Ma age is assigned to Groothoek thrusting (Blignault et al., 1983, p. 22; Van der Merwe and Botha, 1989). C.J.H. Hartnady (personal communication, 1986) suggests that a 1500 Ma age might be appropriate to Groothoek thrusting because he interprets the intensely foliated fabric in the 1500 Ma Eyams granite pluton in the footwall zone of the thrust as forming syn-tectonically. Thus there is considerable controversy with respect to the age of Groothoek thrusting.

In the study area constraints on the age of Groothoek thrusting are provided from field evidence. One small undeformed granitic pluton just west of Kliphoogte, related to the Kromnek granite, intrudes the hangingwall of the Groothoek Thrust zone. This places an upper limit on the age of Groothoek thrusting as older than at least 1000 Ma (cf. section 3.2.12).

The Namaqua metamorphism shows no abrupt change in gradient across the thrust zone (Chapter 5). Metamorphism therefore affected the area across the thrust zone for some distance north of the thrust zone. This observation supports the proposal that thrusting took place before 1100 Ma. A similar relationship exists at Geselskapbank some 70 km to the east (Moore, 1986; Grütter, 1986).

A 1500 Ma age for the main Groothoek thrusting is possible in view of syn-deformational effects in the Eyams granite. This age may be significant in that it approximates the 1600 Ma Sm/Nd minimum age for amphibolites in the upper Aggeneys sequence just south of the Groothoek Thrust in the Pofadder area (Betton, 1984). In this case Groothoek thrusting could be related to deformation in supracrustals of the Aggeneys Sequence. However, the Eyams granite in the study area occurs some 5 to 10 km south of the Groothoek Thrust zone to be markedly affected by it. It is therefore possible that a tectonic event at ≈1500 Ma, unrelated to the Groothoek Thrusting event, may have produced the fabric in the Eyams Granite

4.6.2.4 The D(N)4 episode of deformation

A prominent cleavage ($S(N)_4$) is manifested across the entire width of the zone separating the Richtersveld from the Bushmanland Subprovinces. It is particularly enhanced in discrete thrusts in, for example, the zone of mafic and ultramafic rocks of the Klipbok complex, in the Chabiesies South Thrust, and the Ratelfontein Thrust sheet. Strain is therefore concentrated mainly in dislocation zones. Microfolds (F(N)4) in the Chabiesies South Thrust are interpreted as structures which developed in response to southward movement (Fig. 4.52).

This tectonic episode is probably related to a pervasive phase of late thrusting (Taaibosmond/Skelmfontein thrusting?) recorded by Van der Merwe (1986), Van Aswegen (1988) and Van der Merwe and Botha (1989), in the Namaqua geotraverse.

4.6.2.5 The D(N)5 episode of deformation

An $S(N)_5$ microfabric formed during a northward-directed extensional phase slightly later than 1000 Ma. The effects of this deformational episode are exhibited by brittle deformation textures such as microfracturing in microcline, and kink zones in plagioclase in, for example, the Sabieboomrante adamellite gneiss (cf. Fig. 3.17). Micro-boudinaging in the Chabiesies South Thrust probably developed during this extensional phase.

Tectonism subsequent to the 1000 to 950 Ma period changed to an east-west oriented extensional tectonic setting when a proto-Atlantic ocean formed at the onset of the Pan-African event (Kröner, 1974). According to this model rifting was accompanied by intrusion of the Gannakouriep Dyke Swarm (at ≈878 Ma), and deposition of the north-trending Gariep basin which included pre dominantly clastic sediments at the base, and chemical sediments and glacial deposits upwards in the sequence. Remnants of mafic lavas occur tectonically emplaced in overlying sequences.

4.6.3 Pan-African deformation (D(P) event between ≈750-450 Ma)

4.6.3.1 The D(P) episode of deformation

The earliest Gariep-associated phase of deformation is interpreted as commencing with a left-lateral, strike-slip movement which was southerly directed. During this tectonic episode movement took place along the Stinkfontein Contact Shears, Kromnek Shear and the Steenbok Shear, resulting in moderate displacements probably of the order of 15 to 20 km.

This shearing in basement rocks took place prior to deposition of the Stinkfontein Formation (Joubert, 1971). Ritter (1980, p. 132) confirms this interpretation by noting that the unconformity cuts across these structures in the Eksteenfontein area farther to the north. The foliation in the shear zones developed during the "Devils Castle Event" (Ritter, 1980, p. 132). A ≈700 Ma age is assigned to the shearing, this being the age of the Klipbokkop pluton which intrudes the Pan-African foliation east of Eksteenfontein, and regarded as a correlative of the Richtersveld Suite (Ritter, 1980, p. 236).

Allsopp et al. (1979), however, designated a resetting age of 700 Ma to this tectonic event, approximating the age of the first Pan-African metamorphic imprint south of Eksteenfontein (Chapter 5).

Significant left-lateral movement took place along the steeply-dipping Steenbok Shear during this episode, to account for the large amount of net-slip associated with this shear. During this episode similar folds (F(P)1) developed in the Sabieboomrante adamellite gneiss immediately adjacent to the Steenbok Shear (Plate 12).

4.6.3.2 The D(P)2 episode of deformation

Easterly to south-easterly directed tectonism followed the D(P)1 episode of deformation, forming steeply dipping shears. These shears are interpreted as "reverse" faults, spaced on average 4 km apart across the area (Fig. 4.53(a)). Evidence for direction of tectonic transport lies in the well-developed north-westerly plunging $L(P)_2$ stretching lineation in all the major Pan-African shear zones. Generally, the relative displacement and deformation intensity increase westwards, confirming original observations made by Allsopp et al. (1979).

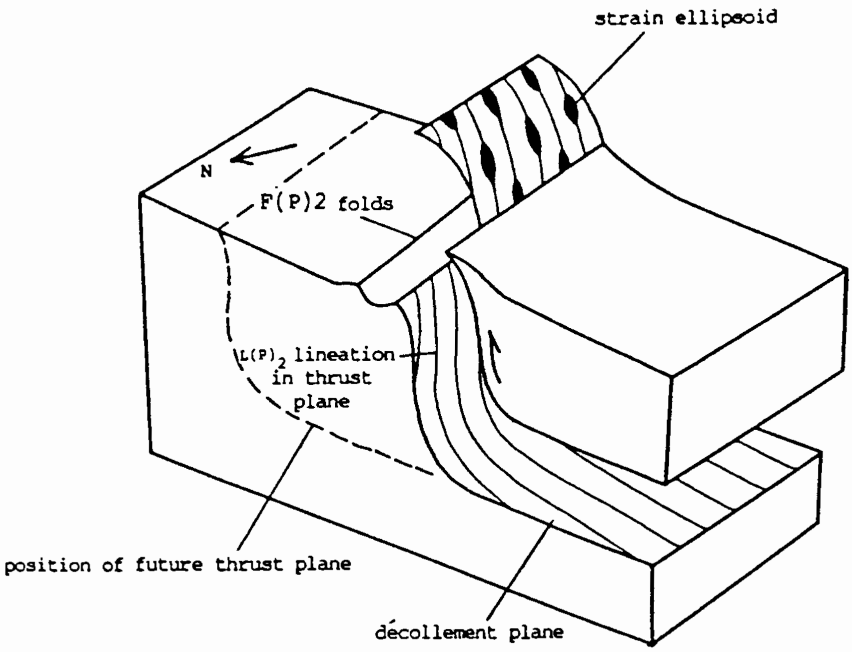
In the extreme west of the area strata of the Stinkfontein Formation at the contact zone show the same $S(P)_2$ and $L(P)_2$ structural imprint as in basement rocks to the east, confirming that tectonism during this episode was post-Stinkfontein Formation in age.

Thrusting during this episode was accompanied by rotation of crustal blocks (Booth, 1988). This developed as the shears were forced to deflect around, or modify their orientation where they transected granitic rock-types, e.g., the Sabieboomrante adamellite gneiss and Kouefontein granite gneiss. The F(P)2 large open fold on Vioolsdrif Toekennings Gebied in the east is interpreted as forming during this stage, in response to a rotational movement on the Chabiesies Shear. Similarly, drag features adjacent to the Steenbok shear, for example, formed during this episode (Fig. 4.29).

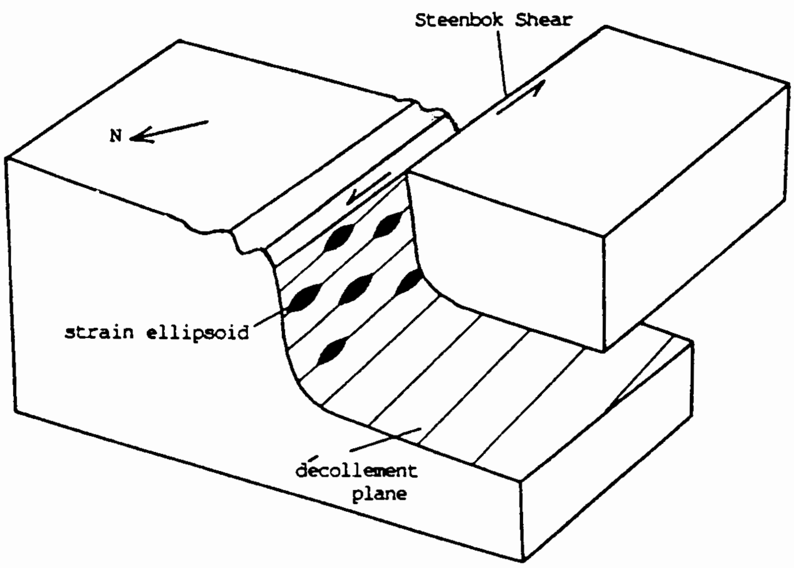
This thrusting episode produced a large net-slip component on Late Proterozoic shears in this area, prior to Nama sedimentation.

4.6.3.3 The D(P)3 episode of deformation

A left lateral, transtensional movement took place along the Steenbok Shear during this episode, after deposition of the Nama Group (Fig. 4.53(b)). During this tectonism a slice of Nama sediments was down-dropped within the Steenbok Shear, in a manner similar to that described by Reading (1980; Fig. 3(a)). The down-dropping of Nama strata is particularly noticeable in the area where there is a change in strike along the Steenbok Shear. This feature conforms with Reading's model where localized zones of extension can occur within a strike-slip fault zone in areas of abrupt deflection in strike orientation, as in the case of the Steenbok Shear. Extensional slip planes in Sabieboomrante adamellite gneiss immediately adjacent to the Steenbok Shear are in-



(a) Development of high- angle thrusts in relation to an inferred décollement plane in depth, during the D(P)2 tectonic episode.



(b) Left-lateral strike-slip movement along the Steenbok Shear, during the D(P)3 tectonic episode.

Fig. 4.53 Schematic block diagrams illustrating movement along steeply-dipping shears in relation to an inferred décollement plane, during the Late Proterozoic - Early Palaeozoic.

terpreted as forming during this deformational episode (Plate 12).

Deformation took place under greenschist facies conditions (Chapter 5), developing muscovite cleavages in the Steenbok Shear (Table 4.1). Down-dropped slices of Nama strata within the Steenbok Shear were deformed into steeply-plunging open and isoclinal folds (F(P)3 folds; Fig. 4.15). Deformation occurred at a slightly higher crustal level than previously, resulting in the development of phyllonites in the Steenbok Shear.

Rotational characteristics of faults are particularly evident in the Kouefontein Shear where it acquires an en echelon form upon transecting the Kouefontein granite gneiss and Sabieboomrante adamellite gneiss (cf. Fig. 4.28). A minor thrust ramp, the Kouefontein North Ramp, probably formed in the crustal block between the Steenbok and Kouefontein Shears, in response to rotation in the latter shear (cf. Figs. 4.46 and 4.47).

Gentle and open folds of F(P)3 generation, with fold axes plunging northwards, formed in regions immediately west of the north-northeasterly striking, steeply-dipping thrust zones (cf. Fig. 4.2(d)). This is particularly evident in crustal blocks east of Steenbok Shear, although occasional F(P)3 open to gentle folds are also present in the western sector, e.g. Kabies se Berg synform (cf. Fig. 4.1).

The age of deformation is younger than 600 Ma, because Gannakouriep dykes and Nama sediments in the Steenbok Shear have been deformed into folds of F(P)3 generation. This deformation phase is probably associated with the 500 Ma resetting event, equating with the "Blackface Mountain mylonite belt" (Ritter, 1980, p. 142).

4.6.3.4 The D(P)4 episode of deformation

The last deformational phase was east-southeasterly directed, forming a cross-cutting $S(P)_4$ cleavage in phyllonites of the Steenbok Shear (cf. Fig. 4.20), but also in strata west of the shear (cf. Fig. 3.10). F(P)4 kink folds formed through shearing during this last easterly-directed episode of deformation, their fold axes having an approximately horizontal orientation (Fig. 4.16).

4.6.3.5 Summary and conclusions

In summary, the D(P) tectonic event comprises four episodes which are strike-slip and thrust related. Tectonism was initially directed towards the south, forming strike-slip faults along the Stinkfontein Contact, Kromnek and Steenbok Shears. This was followed by south-easterly directed tectonism, resulting in the development of high-angle reverse faults (thrusts) with a variable rotational component. The faults segment the western coastal belt into north-northeasterly trending crustal blocks averaging 4 km wide.

Subsequent tectonism during the late Pan-African event commenced with a left-lateral, transtensional movement along the Steenbok Shear in post-Nama times. This movement down-dropped Nama strata within the Steenbok shear and rotated them down adjacent to the Kouefontein Shear. Tectonism was south-easterly directed, forming northward-plunging open folds, but deformation was greater along the Steenbok Shear. The last phase of tectonism was easterly-directed, forming brittle deformation structures in the Steenbok Shear.

From evidence yielded in the study area, the writer concludes that during the Mid-Proterozoic (Namaqua) and Late Proterozoic-Early Palaeozoic (Pan-African) events deformation commenced in the ductile portion of the earth's crust. Regional uplift took place subsequently during each event, and deformation continued at progressively shallower levels until cessation in the brittle zone.

5 METAMORPHISM

5.1 GENERAL ASPECTS

The first broad framework for understanding the regional metamorphic pattern in Namaqualand is that provided by Joubert (1971). He concluded that a west to east prograde metamorphic zonation pattern is present in the region extending from Bitterfontein in the south to the present area. The metamorphic zonation pattern commences with greenschist facies at the coast, increasing to granulite facies in the Springbok area. Joubert (1971) correlates separate metamorphic episodes with deformation episodes in his first event which he interprets as commencing at 2500 Ma, and lasting until 1000 Ma. Amphibolite and greenschist facies metamorphism are associated with shearing during two succeeding, unrelated events, between 1000 and 850 Ma (Joubert, 1971, p. 10).

This general pattern was confirmed by Jack (1980) from an area south-west of Joubert's, between the coast and Garies. However, the grade of metamorphism at the coast corresponds to the epidote amphibolite facies, increasing eastwards to granulite facies (two pyroxene zone). Greenschist facies does not form part of the prograde sequence here, but occurs as a retrograde event in Late Proterozoic shear zones.

Ritter (1980) notes a pattern of metamorphic grade increasing from greenschist facies at the Orange River southwards to merge with Joubert's (1971) high grade zone. Ritter (1980) concludes that an early high-grade metamorphic event occurred between 2000 - 1800 Ma in the south-east Richtersveld, implying that the Namaqua metamorphic event took place at this stage. This conclusion is based on textural evidence showing the former presence of sillimanite in metapelitic rocks. Essentially, Ritter envisages that the Richtersveld rocks and Namaqua gneisses form a continuous succession, having undergone the same tectonothermal history.

Ward (1977, p. 36) examined parageneses and stability ranges of several mineral species in the area immediately to the east of the present area, recording ubiquitous sillimanite in the form of fibrolite in the Namaqualand aluminous schists and gneisses. Mineral species such as cordierite, anthophyllite, staurolite, andalusite and garnet in pelitic rocks are interpreted as showing that a tectonothermal event of regional extent occurred after the emplacement of the Vioolsdrif Suite, and took place under amphibolite facies of metamorphism.

In the Namaqua geotraverse between Vioolsdrif and Springbok, Blignault et al. (1983) record an increase in metamorphic grade from greenschist facies at Vioolsdrif to granulite facies at Springbok. Regionally, each facies of metamorphism reflects tectonically juxtaposed crustal levels, implying that metamorphic discontinuities broadly correspond to tectonic breaks. A prograde metamorphic event is associated with Groothoek thrusting during the Namaqua event (1200-1100 Ma). This metamorphic event overprints the gneisses of the Bushmanland Subprovince, relict granulites being recorded in amphibolite facies rocks of the Steinkopf Terrane (Van Aswegen, 1988). Retrograde metamorphism is associated with medium-grade Groothoek schists in the form of large muscovite porphyroblast development.

Blignault's (1977) conclusions from an area in Namibia extending north-eastwards from the Orange River for some 200 km, are that a main metamorphic event (M1) increases in grade towards the north-east, but decreases again beyond a central high-grade core near the Lord Hill mylonite belt. M1 here is interpreted as commencing as early as 1900 Ma (Blignault, 1977).

Waters (1986) defines a symmetrical pattern of east-west trending metamorphic zones in western Namaqualand, a high-grade granulite facies core centred on the Garies - Kliprand area, some 100 km south of Springbok. The metamorphic history here is explained by intrusion of large volumes of acid magma during conditions of increasing pressure, during the Namaqua event (1200 - 1100 Ma).

Clifford et al. (1981) deduce a ≈1200 Ma age for the Namaqua metamorphism in the Okiep area, based on a Rb/Sr isochron through Nababeep gneiss and quartzo-feldspathic granulite samples in that area. Although this date essentially reflects metamorphic homogenization at 1200 Ma, most investigators accept that it represents the age of the main metamorphic event in the Namaqua mobile belt.

From a study of pelitic rocks (possibly basal Gariep) on Zoutpan, north of Hondeklipbaai, Joubert and Waters (1980) conclude that a metamorphic event later than the high grade metamorphism in gneisses of the Bushmanland Subprovince, but prior to the Pan-African event, formed kyanite and staurolite in the schistose rocks there, and east of Port Nolloth. Their reasoning is based on the observation that outliers of Nama sediments in the area between Port Nolloth and Garies have not been subjected to amphibolite facies metamorphism, although they are situated in the region affected by Pan-African metamorphism.

By recognizing a metamorphic event younger than the Namaqua metamorphic cycle (possibly at ≈700 Ma), Joubert's (1971) original interpretation of greenschist facies metamorphism at the coast is now reinterpreted by Joubert and Waters (1980) as an overprinting event reaching lower amphibolite facies metamorphism, the grade increasing towards the west. In the coastal region south-west of Bitterfontein, Waters, Joubert and Moore (1983) show that two metamorphic events are superimposed on granulite facies basement gneisses and cover rocks in this area. The earlier event occurred at ≈700 Ma, reaching lower amphibolite grade, and is associated with left-lateral rotation of basement structures. The grade of metamorphism increases towards the west, similar to the pattern in the Hondeklipbaai area some 120 km to the north-west of Bitterfontein. The younger event is post- Nama (≈500 Ma), reaching upper greenschist facies metamorphism.

Jackson and Zelt (1984) relate metamorphism and shearing along the west coast to initial phases of Pan-African extensional tectonism and sedimentation.

Pelitic samples from the Stinkfontein Formation between Port Nolloth and Eksteenfontein show that metamorphism associated with the Gariep Belt reached the biotite zone of low grade metamorphism (Ritter, 1980, p. 202). This observation is confirmed by M.W. von Veh (personal communication, 1986) for the Sendelingsdrif area in the northern Richtersveld.

A summary of regional metamorphic patterns and metamorphic development in the Richtersveld and Bushmanland Subprovinces, as outlined by the above-mentioned authors, is given as follows:

(i) The Richtersveld Subprovince is a domain of low grade greenschist facies metamorphism from where the grade increases outwards towards the north, east and south (cf. Fig. 2.3). Deformation and metamorphic events and episodes are interpreted by Joubert (1971) and Blignault (1977) as commencing prior to 1900 Ma, and continuing until the close of the Namaqua event at ≈1000 Ma. Major thrusting along the southern boundary of the subprovince took place just after 1100 Ma, i.e., during the latter part of the Namaqua event (Blignault et al., 1983). The regional metamorphic imprint in the Richtersveld Subprovince is therefore interpreted as commencing during the Eburnian period (Blignault, 1977; Ritter, 1980);

(ii) Deformation and metamorphism in the Bushmanland Subprovince is interpreted by Clifford et al. (1981) and Waters (1986) as having taken place during the Namaqua event at 1200 - 1100 Ma. The metamorphic pattern in this subprovince shows an increase in metamorphic grade southwards from the Richtersveld Subprovince towards a granulite facies core centred on Garies-Kliprand, some 100 km south of Springbok. At its southern marginal zone the Richtersveld Subprovince has been thrust southwards over the Bushmanland Subprovince along the Groothoek Thrust zone. There remains a major controversy as to whether metamorphism is continuous from the Richtersveld Subprovince into the Bushmanland Subprovince. Did this metamorphism commence as early as 1900 Ma, as proposed by Joubert (1971), or has the Namaqua metamorphic event overprinted the Early Proterozoic event of the Richtersveld Subprovince, as proposed by Clifford et al. (1983)? In the latter case earlier (Eburnian ?) metamorphic imprints in the Bushmanland Subprovince would have been obliterated. It is also not clear whether over-thrusting along the

Groothoek Thrust zone took place during an earlier 1800 Ma period as proposed by Moore (1986), or subsequent to the Namaqua event (Blignault et al., 1983). The metamorphic pattern across the Groothoek Thrust zone is therefore of major significance in elucidating the development of thrusting in relation to metamorphism along this significant tectonic lineament;

(iii) Along the west coast a post-Namaqua metamorphic event has affected both the Gariep cover rocks and gneisses of the Bushmanland Subprovince. This event reached amphibolite facies conditions in the Hondeklipbaai area and near the coast to the west of Bitterfontein, some 200 km south of the study area. It is significant to determine whether there is any evidence for this metamorphic event in the present study area. Metamorphism in the Gariep Belt reached greenschist facies metamorphism (Ritter, 1980), but it is not clear how this is related to Pan-African tectonism in the region;

(iv) A post-Nama metamorphic event (at ≈500 Ma) reached upper greenschist facies metamorphism in the area south-west of Bitterfontein. It has yet to be established how far northwards this metamorphic event has made its imprint.

5.2 METAMORPHIC CONDITIONS WITHIN THE STUDY AREA AS REVEALED THROUGH DIAGNOSTIC ROCK-TYPES

Rocks most useful for elucidating metamorphic conditions in the study area are aluminous meta-pelites and meta-basites. Metamorphosed calcareous rocks which, together with the latter two rock-types, could serve as sensitive indicators of temperature and pressure conditions, are absent from this region. However, textural evidence from metavolcanic and ultramafic rocks yields additional useful information on conditions of metamorphism.

Metamorphic textures may reveal the degree to which equilibrium was attained during metamorphism. Mineral assemblages and their textural relationships in turn are products of reactions, during which changes are brought about by the intensive variables temperature, pressure and composition of the fluid phase. These changes cause diffusion and migration of chemical species, and nucleation and growth of new minerals resulting in an overall lowering of the free energy of the system.

A study of metamorphic rocks in thin section helps with the identification of index minerals. However, only a set of coexisting minerals which defines a metamorphic assemblage is a potential petrogenetic indicator (Winkler, 1974, p. 27).

The term "grade", refers to the intensity of metamorphism, and as such is a function of all three intensive variables, temperature, pressure and composition of the fluid phase. Grade can be correlated with the various metamorphic facies, as in Table 5.1.

TABLE 5.1 Correlation of metamorphic grade with facies		
Winkler, (1974, p.7)	Best (1982, p. 372)	Miyashiro (1973, p. 66)
very low grade (190° - 360°C)	zeolite facies	
low grade (360° - 500°C)	greenschist facies	greenschist facies
medium grade (500° - 600°C)	lower amphibolite facies	epidote-amphibolite facies
high grade (600°C --->)	upper amphibolite facies	amphibolite facies
high grade (regional hypersthene)	granulite facies	granulite facies

The study of microtextures can give useful information about the timing of metamorphism relative to deformation (Zwart, 1960; Johnson, 1963; Rast, 1962; 1964; Spry, 1969; Ghosh, 1975). In particular, external fabric (Se) relative to inclusion trails (Si) can provide crucial information for deciding whether a particular porphyroblast is pre-, syn- or post-tectonic (terminology after Sander, in Whitten, 1966, p. 481).

In cases where the foliation is "bowed-out" around a porphyroblast, controversial ideas are put forward to explain the development of this texture. Some authors suggest that the "bowing-out" of the foliation is due to a pushing aside effect caused by a "crystallization force", during growth of the porphyroblast (Misch, 1971; Saggerson, 1974), whereas others ascribe this texture to tectonic flattening after development of the denser porphyroblasts (Shelley, 1972; Vernon and Powell, 1975). The latter authors demonstrate that there are cases where several episodes of movement and crystallization history are determined from the growth of a particular porphyroblast.

Spry (1969) interprets the "bowed out" foliation effect as resulting from dissolution and mass transfer of certain minerals in the matrix (for growth of the porphyroblasts), or by pressure solution (to produce flattening).

A careful description and interpretation of metamorphic textures is therefore essential when relating metamorphism to deformation. Interpretations of the development of the above-mentioned texture would more than likely favour a tectonic origin where rocks containing these textures occur in zones of strong shearing. In an attempt at simplifying interpretations of metamorphic textures in the study area I propose the following:- those textures related to the regional $S(N)_2$ fabric formed during the Namaqua event, whereas textures in, or associated with, northerly-trending shear zones formed during the Pan-African event.

5.2.1 Subdivision of study area into domains

The study area is one that has experienced polymetamorphism. The first recognisable metamorphic event (upper greenschist to upper amphibolite facies) imprinted across the area took place between 1200 - 1100 Ma (Namaqua event). This event obliterated all evidence of prior metamorphism (if any) in the area.

A retrograde event followed shortly after 1100 Ma, evidence for this being recorded only in the eastern part of the study area. West of Tierkloof Shear there is evidence for a lower amphibolite facies imprint which took place during the Pan-African event (700 Ma). This was followed by retrograde metamorphism focussed in northerly-trending Pan-African shear zones, between 500 and 470 Ma (section 5.3).

Because of the complex metamorphic overprinting, the region has been sub-divided into five domains for descriptive purposes (Fig. 5.1). The boundaries of these domains are chosen to correspond to some of the major north-trending Pan-African shear zones. East of Steenbok Shear metamorphic domains do not coincide with structural domains (Chapter 4), but are simplified into two domains, viz. 4 and 5. In this way areas showing relatively homogeneous metamorphic patterns are delineated.

5.2.2 Metamorphism of pelitic rocks

5.2.2.1 Mineral assemblages

5.2.2.1.1 Domain 1 - between Stinkfontein Contact Shears and Kromnek Shear

In the southern portion of Domain 1, just above the basal quartzites of the Ratelfontein suite and within the zone affected by the Stinkfontein Contact Shears, metapelitic rocks have the mineral assemblage

muscovite + biotite + staurolite + garnet + quartz.

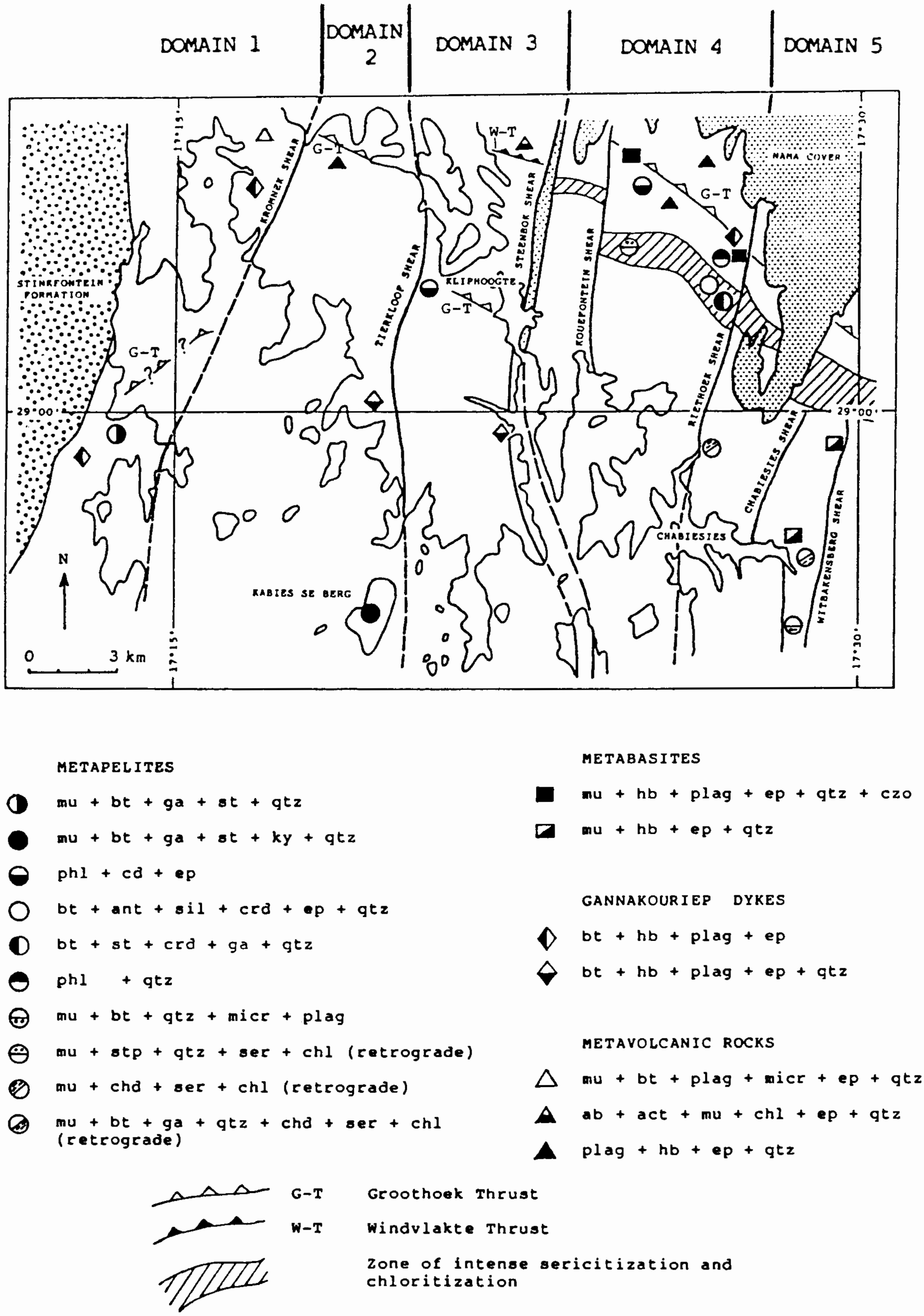


Fig. 5.1 Subdivision of study area into metamorphic domains, as revealed through mineral assemblages of metapelites, metabasites, Gannakouriep dykes and metavolcanic rocks.

Al_2SiO_5 polymorphs are not present, although kyanite is recorded 15 km to the south-west on Oograbies West (Joubert, 1971), and in shear zones farther north in the Richtersveld (De Villiers and Söhnge, 1959; Ritter, 1980). Textural relationships in metapelites of the Ratelfontein suite show that chloritoid is completely replaced by chlorite, having initially developed as large euhedral to subhedral porphyroblasts, and still retaining some of its characteristic physical features (cf. Fig. 3.10, and section 3.3.1.2). Brown biotite, staurolite and garnet form porphyroblasts which have grown across the $S(P)_2$ fabric (Fig. 5.2(a)). The original assemblage therefore comprises chloritoid, biotite, garnet and staurolite, together with muscovite and quartz (Fig. 5.3(a)), chloritoid being replaced by chlorite during a subsequent retrograde event.

Many staurolite porphyroblasts show one or two euhedral margins, whereas opposite margins form digitations mimetically developing parallel to $S(P)_2$, outlined by thin needles of muscovite (Fig. 5.2(a). Some large staurolite porphyroblasts show syn-deformational textures, quartz inclusions clearly defining inclusion trails (Si) deformed into an open S-shape (Fig. 5.2(b)). I interpret the latter texture as forming during the early stages of the Pan-African event (possibly soon after the D(P)1 tectonic episode). Porphyroblasts mostly formed interkinematically, being only slightly affected by a later buckling of the foliation (Fig. 5.4).

Staurolite could have formed by either of the following two discontinuous reactions:

(i) in the presence of chlorite

$$\text{chloritoid} + \text{chlorite} + \text{muscovite} \rightleftharpoons \text{staurolite} + \text{biotite} + \text{quartz} + H_2O \qquad (1)$$

(Baltatzis, 1979);

(ii) after chlorite has reacted out chloritoid finally disappears by the terminal reaction

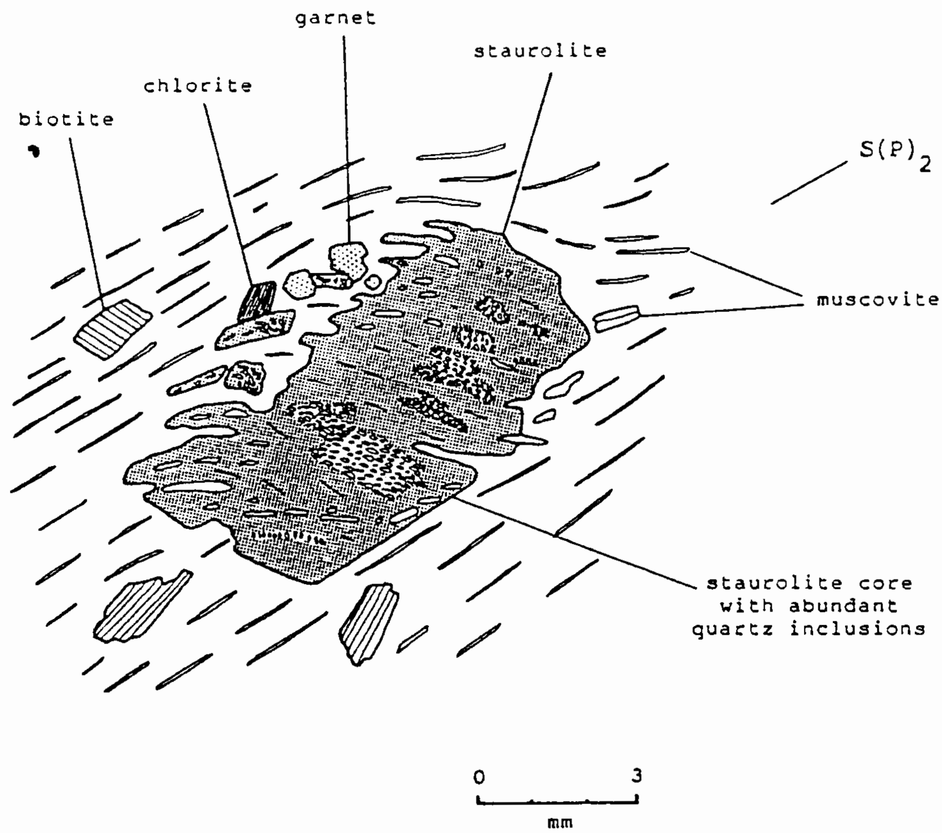
$$\text{chloritoid} + \text{muscovite} + \text{quartz} \rightleftharpoons \text{staurolite} + \text{biotite} + \text{garnet} + H_2O \qquad (2)$$

(Baltatzis, 1979).

Both reactions mark the transition from greenschist to amphibolite facies, reaction (1) at somewhat lower temperature than (2). The assemblage in central Domain 1 (Fig. 5.3(a)) probably indicates equilibrium conditions for reaction 2.

This assemblage places additional constraints on the pressure. As both reactions are located below the invariant point chloritoid + staurolite + garnet + biotite + chlorite, pressure must be less

(a)



(b)

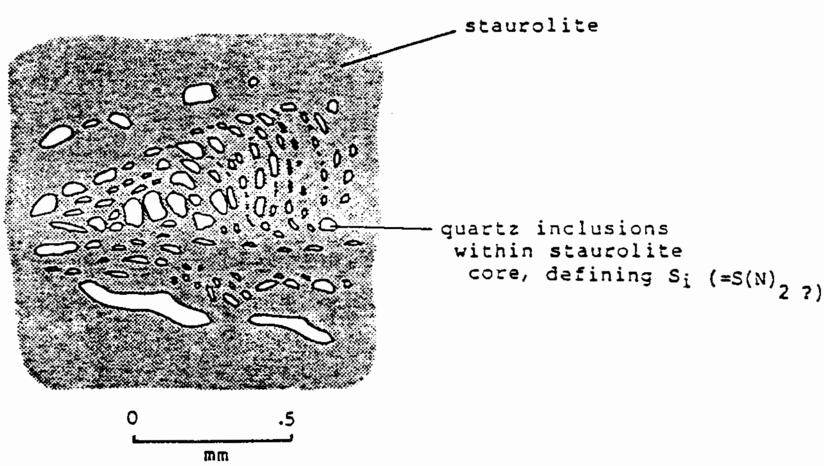
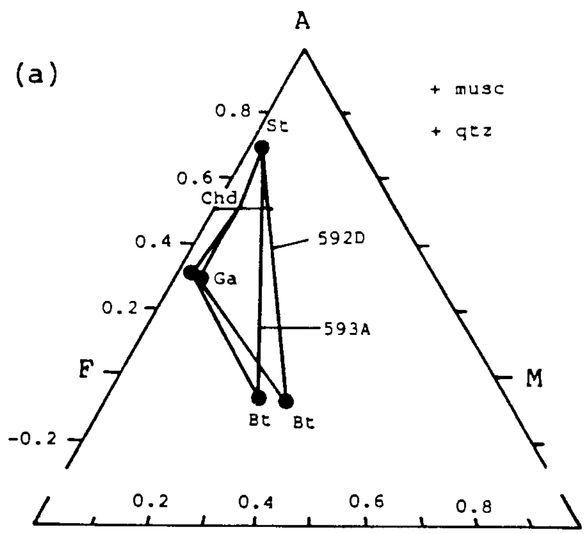


Fig. 5.2 (a) Large staurolite porphyroblast with quartz-rich cores from a metapelite, Ratelfontein suite, near Penssleep beacon, Domain 1.

(b) Syn-deformation texture shown by Si within quartz-rich core of staurolite porphyroblast.

(a)



(b)

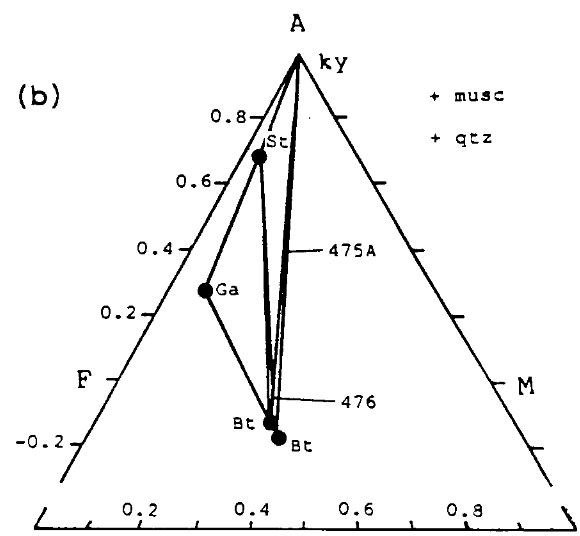


Fig. 5.3 A F M projections from muscovite (after Thompson, 1957), illustrating compositions of coexisting minerals in metapelites.
A: $(Al_2O_3 - 3K_2O)/ (Al_2O_3 - 3K_2O + MgO + FeO)$.
M: $MgO/(MgO + FeO)$.
(a) Samples 592D and 593A from the Ratelfontein suite, Domain 1 (Pan-African event). Note that average compositions of staurolite, garnet and biotite are plotted whereas chloritoid is drawn schematically.
(b) Samples 475A and 476 from the Chabiesies suite at Kabies se Berg, Domain 2 (Pan-African event). Note that average compositions of staurolite, garnet and biotite are plotted.

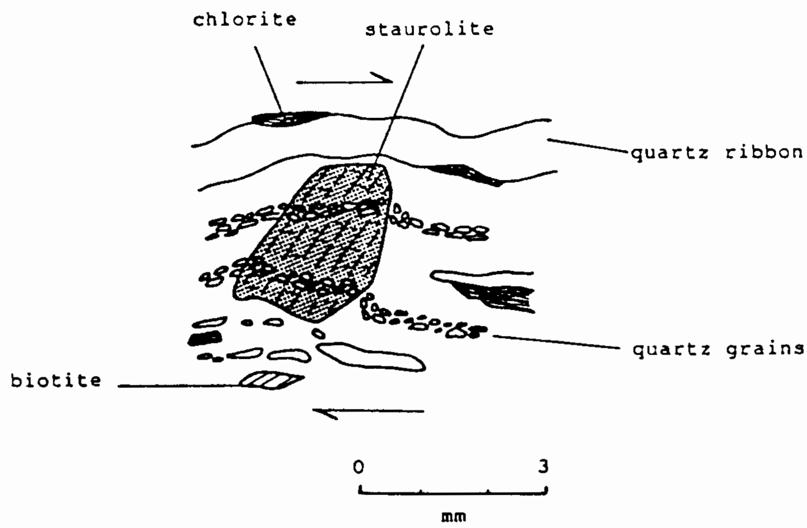


Fig. 5.4 Post-tectonic growth of staurolite porphyroblast, and subsequent slight buckling. Sample from metapelite Ratelfontein suite, near Penssleep beacon, Domain 1.

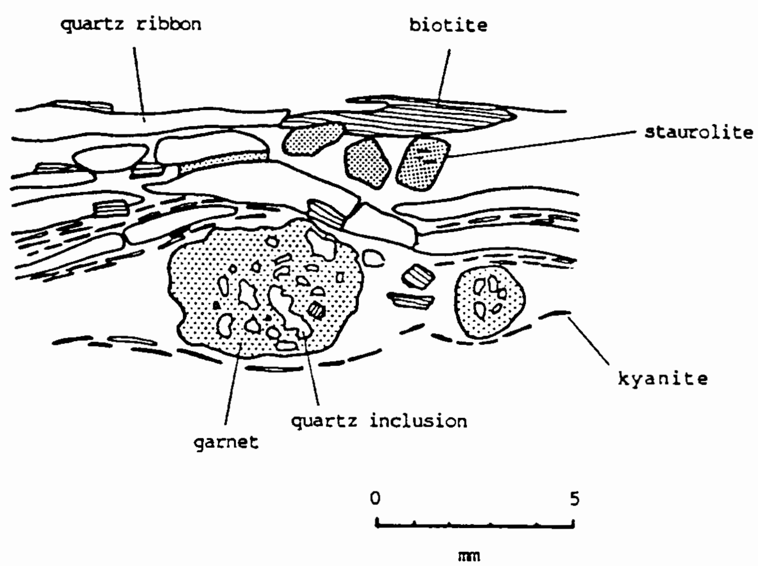


Fig. 5.5 Oval and circular shaped garnet porphyroblasts between a foliation defined by quartz ribbons, kyanite blades and phyllosilicates, from a mylonitized metapelite bed, Kabies se Berg.

than 6 kbar (Harte, 1975). The presence of kyanite 15 km to the south-west on Oograbies West suggests that the pressure is greater than 4.5 kbar.

5.2.2.1.2 Domain 2 - between Kromnek and Tierkloof Shears

At Kabies se Berg in the extreme south the following mineral assemblage is present,

muscovite + biotite + garnet + staurolite + kyanite + quartz + plagioclase

Chlorite replaces biotite grains along their edges and is considered to be a retrograde product on biotite. The original mineral assemblage is therefore represented on an AFM diagram as in Fig. 5.3(b)). This assemblage is typical for the middle amphibolite facies and indicates that the metamorphic grade is slightly higher than that in Domain 1. The presence of garnet, staurolite, kyanite and biotite suggest that these phases grew in aluminous- rich portions of the metapelite, at slightly greater pressure conditions than assemblages in Domain 1.

Kyanite in this assemblage may have formed either by the continuous reaction

$$\text{staurolite + muscovite + quartz} \Leftarrow\Rightarrow \text{kyanite + biotite} + H_2O \qquad (3)$$

or by the discontinuous reaction

$$\text{staurolite + chlorite + muscovite + quartz} \Leftarrow\Rightarrow \text{kyanite + biotite} + H_2O. \qquad (4)$$

Textural development of garnet porphyroblasts in metapelite units from Kabies se Berg shows oval- shaped porphyroblasts containing few randomly oriented quartz inclusions (Fig. 5.5). The oval shape suggests that the garnet porphyroblasts may have formed through a deformation partitioning mechanism similar to that described by Bell et al., (1986). On the other hand, faint spiral arrangement of quartz grains in the garnets suggests that the garnet may have experienced at least some rotation during its formation. Both mechanisms are, however, indicative of syntectonic development of garnet in the Chabiesies suite metapelites at Kabies se Berg.

In contrast to the large staurolite porphyroblasts in Domain 1 staurolites in Domain 2 are usually small, euhedral, prismatic grains.

5.2.2.1.3 Domain 3 - between Tierkloof and Steenbok Shears

No metapelites containing minerals diagnostic for temperature and pressure estimates were located in this domain.

5.2.2.1.4 Domain 4 - between Steenbok and Riethoek Shears

In this domain and those farther to the east remnants of the effects of the Namaqua metamorphic assemblages (1200 - 1100 Ma event) are recorded. Metapelitic rocks are characterized by replacement textures. Sericite and chlorite completely pseudomorph certain minerals, whereas they replace others to varying degrees. Reliance is placed on crystallographic shape, habit and other characteristics of the pseudomorphed minerals in order to determine original mineral assemblages.

By stripping off the effects of retrograde metamorphism, the following mineral assemblages represent original parageneses (Namaqua age) in metapelitic and semi-pelitic rocks of the Groenrivier suite:

(i) phlogopite + corundum + muscovite + epidote

In this assemblage large porphyroblasts of corundum are replaced to varying degrees by sericite (cf. Fig. 3.12). Phlogopite shows preferred growth parallel to $S(N)_2$;

(ii) cordierite + anthophyllite + sillimanite + biotite + quartz

This assemblage is extremely localized in the present area, being recorded only in very thin units near the base of the Ratelfontein suite. Lal and Shukla (1975) propose that such an assemblage is likely to have formed within a temperature range of 500°-600°C, and between 3 - 5 kbar pressure.

Anthophyllite here could have formed from the reaction

$$9 \text{ talc} + 4 \text{ forsterite} \rightleftharpoons 5 \text{ anthophyllite} + 4 H_2O \qquad (5)$$

(Winkler, 1967).

Winkler (1967) suggests that such a reaction takes place in the high temperature range of the hornblende-hornfels facies. Investigations by Greenwood (1963) of the stability of pure magnesium anthophyllite indicate a lower stability limit of 667° ± 8°C. Anthophyllite in the assemblage above may therefore have formed at temperatures slightly higher than those proposed by Lal and Shukla (1975);

(iii) biotite + staurolite + cordierite + garnet + quartz

This assemblage occurs in metapelites of the Ratelfontein suite which here coincide with a north-westerly trending zone of retrograde metamorphism (cf. Fig. 5.1). Prograde mineral phases plotted on an AFM diagram (Fig. 5.6) show that mineral compositions of staurolite, garnet and biotite in Domain 4 differ slightly from those in Domains 1 and 2 (cf. Fig. 5.3). Biotites from Domain 4 metapelites are notably richer in Mg compared to those of other domains.

Textural relationships in this metapelite (Fig. 5.7) reveal important aspects of its crystallization history. Biotite is the most abundant mineral, forming broad blades aligned mainly parallel to $S(N)_2$. The staurolite porphyroblasts are surrounded by an inner rim of chlorite and an outer rim of cordierite. Although the cordierite rim is completely pinitized, its original crystallographic shape and splay fractures are still apparent. The staurolite core is extensively sericitized and only a few remnants of staurolite remain. Inclusion trails (Si) of graphitic, or opaques, oriented approximately at right angles to the external foliation (Se), occur in the porphyroblasts. I interpret this texture as post-tectonic, based on the fact that Si has a parallel arrangement and occurs at a steep angle to Se ($S(N)_2$). Textural relationships therefore reveal that staurolite has survived as an armoured relic during metamorphism of the metapelite (Namaqua metamorphic event at 1200 Ma).

The cordierite rim clearly replaces staurolite, indicating a decrease in pressure. This probably reflects the reaction

$$\text{staurolite} + \text{biotite} \rightleftharpoons \text{cordierite} + \text{garnet} + H_2O \qquad (6)$$

in the system $Al_2O_3 - K_2O - FeO - MgO - SiO_2 - H_2O$.

According to Harte (1975) this reaction is pressure dependant and it indicates a decrease in pressure during or after the peak of metamorphism.

A chlorite rim separates the mineral phases staurolite and cordierite (Fig. 5.7). This indicates that chlorite here formed by retrograde reactions after peak conditions were reached during the Namaqua event. Sericite then replaced these minerals.

(iv) muscovite + stilpnomelane + quartz + opaques

This assemblage occurs east of Tierkloof Shear in semi-pelitic rocks of the Ratelfontein suite. The assemblage occurs in the same zone of retrograde metamorphism as in assemblage (iii) above. Stilpnomelane forms large subhedral grains with characteristic golden-yellow to deep reddish-brown pleochroism, in places showing

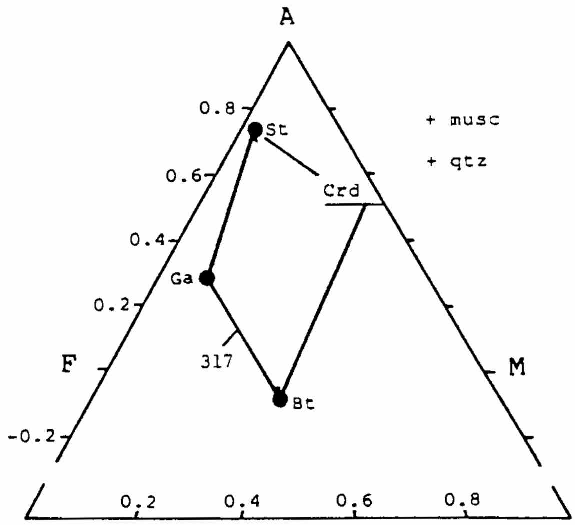


Fig. 5.6 A F M projection from muscovite (after Thompson, 1957) illustrating mineral compositions from metapelite of the Ratelfontein suite, Domain 4 (sample 317). Note that average compositions of staurolite, garnet and biotite are plotted, whereas cordierite is drawn schematically.

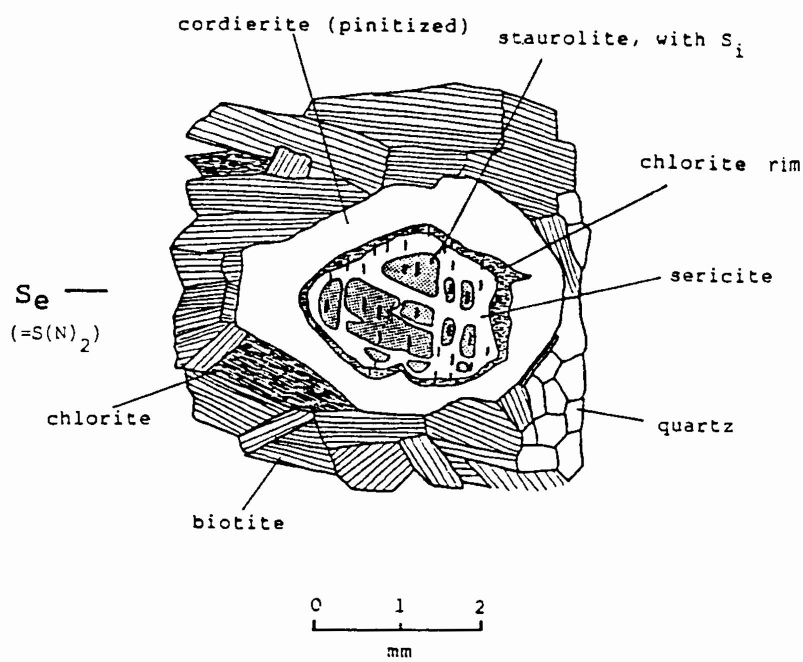


Fig. 5.7 Textural features showing relationship between Si within staurolite grain, and Se. Note cordierite rims staurolite, the former extensively replaced by sericite. A rim of chlorite outlines original size of staurolite porphyroblast.

advanced replacement by chlorite.

According to Heinrich (1965, p. 293) stilpnomelane is relatively common in low grade metamorphic rocks. Miyashiro (1963, p. 263), however, suggests that stilpnomelane is rarely found in medium and low pressure metamorphic rocks, but is more widespread in high pressure metamorphic environments. I interpret stilpnomelane in semipelitic rocks of the Ratelfontein suite as a retrogressive product on high grade rocks (section 5.3).

5.2.2.1.5 Domain 5 - east of Riethoek Shear

Metapelites in the southern portion of this domain have the following mineral assemblages

muscovite + chlorite(penninite) + chloritoid + sericite
muscovite + biotite + garnet + chloritoid + quartz
+ sericite + chlorite
muscovite + biotite + quartz + microcline + plagioclase
+ sericite (after sillimanite)

The presence of microcline (and originally sillimanite) in the latter assemblage shows that muscovite + quartz have reacted to K-feldspar + sillimanite during the peak of metamorphism. I interpret this as good evidence for upper amphibolite facies conditions in the south-east of the area, during the Namaqua event.

Metapelites in the area east of Riethoek Shear generally show advanced replacement by sericite, making the original textural relationships difficult to assess. However, the following observations and interpretations are made from thin sections.
Garnet occurs mostly as remnant grains, surrounded by sericite. This indicates retrogression after the peak metamorphic conditions were reached during the Namaqua event. Sericite, however, occurs mostly in broad bands transecting the rock. This is interpreted as a replacement of sillimanite which originally formed the schistosity. In places sericite retains the needle-like habit of sillimanite, especially in quartz grains in the metapelites. Large muscovite and chlorite porphyroblasts, and small euhedral, randomly-oriented chloritoid grains grow across the matted sericitic texture (cf. Fig. 3.5). I interpret the latter texture as a later prograde development on previously retrogressed high-grade assemblages in the south-eastern part of the study area.

The original mineral assemblages of metapelites in the southern portion of this domain can therefore be represented on an AFM diagram as shown in Fig. 5.8(a). The diagram shows that mineral phases differ from those of other domains, especially biotite which is iron- rich. Retrogressive phases are shown in Fig. 5.8(b).

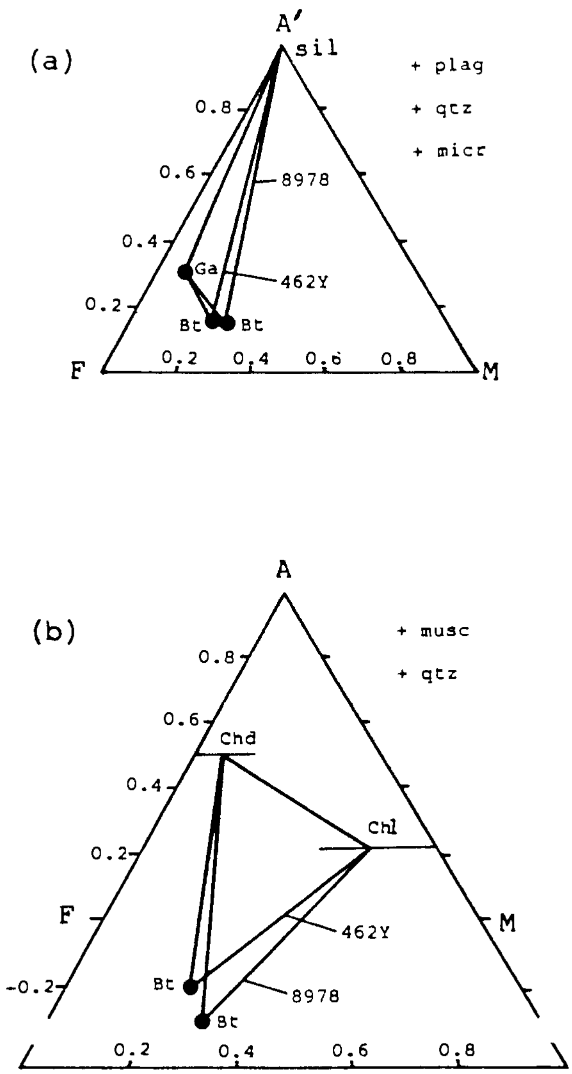


Fig. 5.8 A F M projections (after Thompson, 1957) illustrating mineral compositions in the Chabiesies suite, Domain 5. Samples are numbered 462Y and 8978.

(a) A'F M projection from K-Feldspar, illustrating high-temperature / low pressure assemblages formed during the Namaqua event.
A': $(Al_2O_3 - K_2O)/ (Al_2O_3 - K_2O + FeO + MgO)$.
M : $MgO/ (MgO + FeO)$

(b) A F M projection from muscovite of minerals which have formed on partly retrograded high-grade assemblages. Note that average compositions of biotite are plotted, whereas chloritoid and chlorite are drawn schematically.
A : $(Al_2O_3 - 3K_2O)/ (Al_2O_3 - 3K_2O + FeO + MgO)$.
M : $MgO/ (MgO + FeO)$.

The absence of staurolite suggests that prograde mineral assemblages were probably formed by reactions which surpass staurolite's upper stability limit, e.g.,

staurolite + muscovite + quartz $\rightleftharpoons$
garnet + biotite + Al_2SiO_5 + H_2O (7)
(Carmichael, 1969; Kasch, 1987).

These prograde assemblages show that upper amphibolite facies metamorphism took place in the south east of the area, during the 1200 - 1100 Ma (Namaqua) event.

In localized horizons along the Chabiesies South Thrust zone broad bands of yellowish, very finely matted sericite (after sillimanite) enhance the foliation, and skeletal intergrowths of chlorite and quartz are pseudomorphous after biotite (cf. Plate 20). This texture is very characteristic of some Namaqualand high-grade metapelites showing biotite + quartz (and sometimes sillimanite + quartz) embaying and replacing garnet, for example, and is explained as a back-reaction with liquid in partially melted rocks (D.J. Waters, personal communication, 1986).

Sericite is interpreted as forming predominantly during a period of uplift and cooling, soon after 1100 Ma and associated with a spaced cleavage ($S(N)_4$; Chapter 4). During this retrograde phase garnet and sillimanite were extensively replaced.

Chloritoid and penninite are interpreted as developing during the Pan-African event (700 Ma) because they occur as porphyroblasts across sericitic bands in shear zones such as the Chabiesies South Thrust (see section 5.3.3). They probably formed by nucleation and growth during dehydration reactions from white mica and chlorite.

5.2.2.2 Mineral compositions

5.2.2.2.1 Staurolite

Staurolite-bearing metapelites were collected from the Ratelfontein suite (Domains 1 and 4), and the Chabiesies suite (Domain 2) (cf. Chapter 3). Microprobe analyses were carried out on 18 staurolite grains in meta-pelitic rocks of all three domains (Appendix C-1). Major and trace elements which show up the differences between staurolites of the different domains include:

(i) Al_2O_3 - Staurolites in Domains 1 and 2 have a slightly lower content (53,4 wt%) than those from Domain 4 (55,0 wt%);

(ii) FeO and MgO - Both are slightly higher in Domains 1 and 2 than in Domain 4 (FeO by about 2,0 wt%. MgO by about 0,5 wt%);

(iii) TiO_2 - This is slightly higher in staurolites from Domain 4 (0,56 wt%) compared to those of Domains 1 and 2 (ave. 0,45 wt%);

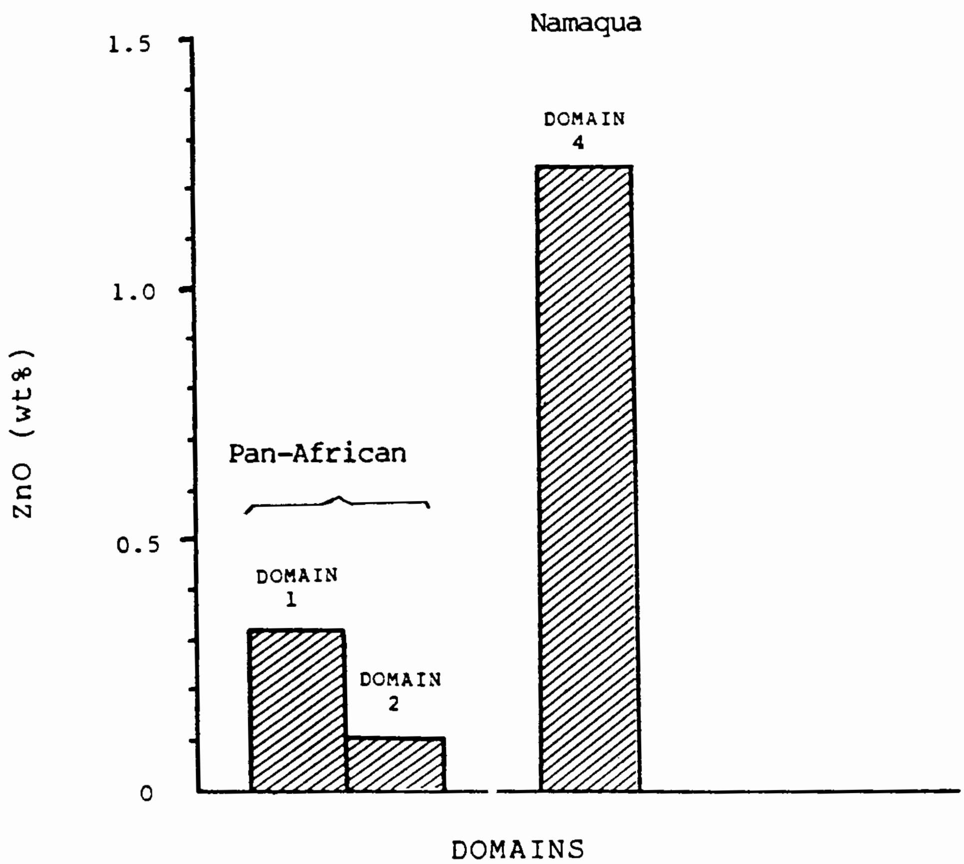
(iv) ZnO - Staurolites from Domain 4 contain notably greater proportions of ZnO (mean 1,24 wt%) than those from Domains 1 and 2 (Fig. 5.9). A plot of ZnO vs $FeO/FeO+Al_2O_3$ demarcates staurolites into three separate groups (Fig. 5.10). Staurolites of Domain 1 have three times the proportion of ZnO (0,32 wt %) compared to those of Domain 2 (0,10 wt %). This discrepancy is primarily a reflection of staurolite occurrence in metapelite horizons of the Ratelfontein (Domain 1) and Chabiesies (Domain 2) suites which contain different mineral assemblages and therefore have different chemical compositions.

From the above observations it appears that the geochemistry of staurolite agrees with textural observations and interpretations. The staurolites of Domain 4 are of an earlier generation (Namaqua event) whereas those of Domain 1 and 2 formed during the Pan-African event, reaching lower amphibolite facies in the West Coast Belt (Section 5.3).

The significance of zinc in staurolite is that its abundance may reveal information on grade of metamorphism and possible genesis of the staurolite. Staurolite has a complex structure, consisting essentially of alternating layers of kyanite and Fe Al $(OH)_2$, arranged parallel to the 010 crystallographic face, thus giving it a good cleavage (Smith, 1968; Griffen and Ribbe, 1973). It has the generalised formula $Al_{18}Fe_4Si_8O_{48}H_4$, extensive cation substitutions taking place in tetrahedral and octahedral sites (Smith, 1968). The principal substitutions are $Al \Leftrightarrow (Fe,Mg)$ in the Al^{6+} site and $Fe \Leftrightarrow (Zn, Al)$ in the Fe^{4+} site, but Zn is also present in the Al^{6+} site and Mg in the Fe^{4+} site (Griffen and Ribbe, 1973). Staurolite accommodates most of the Zn available in metapelites of middle amphibolite grade at the expense of other elements, showing a preference for tetrahedral sites (Miyake, 1985).

The greater abundance of ZnO in Domain 4 staurolites compared to Domains 1 and 2 indicates that the temperature attained was slightly higher in the former than in the latter two domains. This is supported by mineral assemblages in metapelites of the Ratelfontein suite which show that at least middle amphibolite facies was reached in the eastern part of the study area. This pattern suggests ZnO in staurolite increased with grade increasing concentration taking place with decrease in modal proportions in the rock (Guidotti, 1970). Furthermore Zn probably became concentrated in staurolite and was immobile during middle amphibolite facies metamorphism (Tuisku et al., 1987).

Staurolites with high ZnO content are thought to form from desulphidization of sphalerite during metamorphism, as in the Bleikvassli deposit, Norway, where staurolites contain up to 8.77 wt%



Domain	Analyses	Mean	Std. dev.
1	5	0.32	0.08
2	6	0.10	0.05
4	7	1.24	0.14

Fig. 5.9 ZnO content in staurolites from Domains 1, 2, and 4.

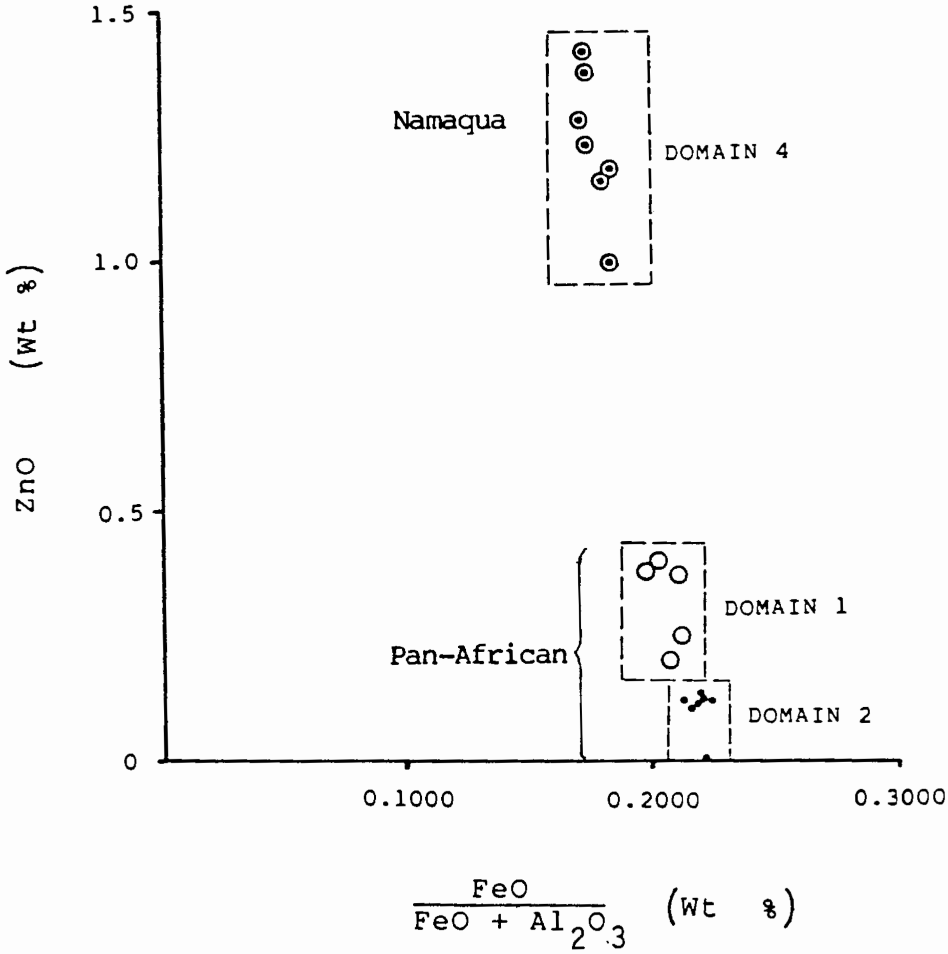


Fig. 5.10 Plot of ZnO vs $FeO/FeO+Al_2O_3$, from staurolites in Domains 1, 2, & 4.

ZnO (Spry and Scott, 1986). Moore and Reid (1989) propose this mechanism for the formation of staurolite at Kraaifontein, some 120 km southeast of the study area. Where non-desulphidization mechanisms are involved in the formation of staurolite ZnO content is usually less than 6 wt % (Spry and Scott, 1986). Staurolites in all three domains in the study area are not associated with any known sulphide occurrences and contain far less than the latter proportions of ZnO. They were probably formed by non- desulphidization mechanisms during amphibolite facies metamorphism.

5.2.2.2.2 Garnet

Core and rim analyses were carried out with the microprobe on 34 garnets from metapelites of the Ratelfontein and Chabiesies suites in Domains 1, 2, 4, and 5, to determine the extent of zoning (Fig. 5.11). Traverses were run across four garnets, one from each domain, to reveal detailed zoning profiles (Fig. 5.12).

Observations

(i) Small garnets (0,05 mm) are as strongly zoned as large ones (12,0 mm), e.g., compare scales of Fig. 5.12(a), (b), (c) and (d).

(ii) Garnets from the Ratelfontein suite, Domain 1, show a variable but general increase in FeO, MgO from core to rim, and decrease in MnO (Fig. 5.13). FeO, however, shows an erratic range of values in the profile of a traversed garnet grain (Fig. 5.12). This can be attributed to analyses made in proximity to quartz inclusions, resulting in an irregular zoning profile (Yardley, 1977). Although MnO concentrations in this garnet vary only slightly (between 0,10 and 0,23%), its profile does have a subdued bell-shape. CaO shows a variable but general increase from core to rim (Fig. 5.13).

(iii) Garnets from the Chabiesies suite in Domain 2 show slight increase in FeO and MnO from core to rim, but decrease in MgO (Figs. 5.12(b), and 5.14). CaO shows a slight increase towards the rims. Occasional high FeO values occur in the centre of the garnet. This can be explained by proximity of quartz inclusions where above normal concentrations of Fe are present. MgO shows virtually no variation across most of the core of the grain, but decreases towards the margins.

(iv) Garnets from the Ratelfontein suite, Domain 4 show a decrease in MnO, FeO from core to rim , but increase in MgO (Figs. 5.12(c), and 5.15). CaO shows no variation at all.

(v) Profiles of garnets from the Chabiesies suite in Domain 5 (samples 462, 8978, Fig. 5.16) have patterns different from those of other garnets in the area. Mn concentration increases dramatically towards the rims of the garnet (Figs. 5.12(d), and 5.16). This is offset by decreasing MgO and FeO towards the rims. CaO shows

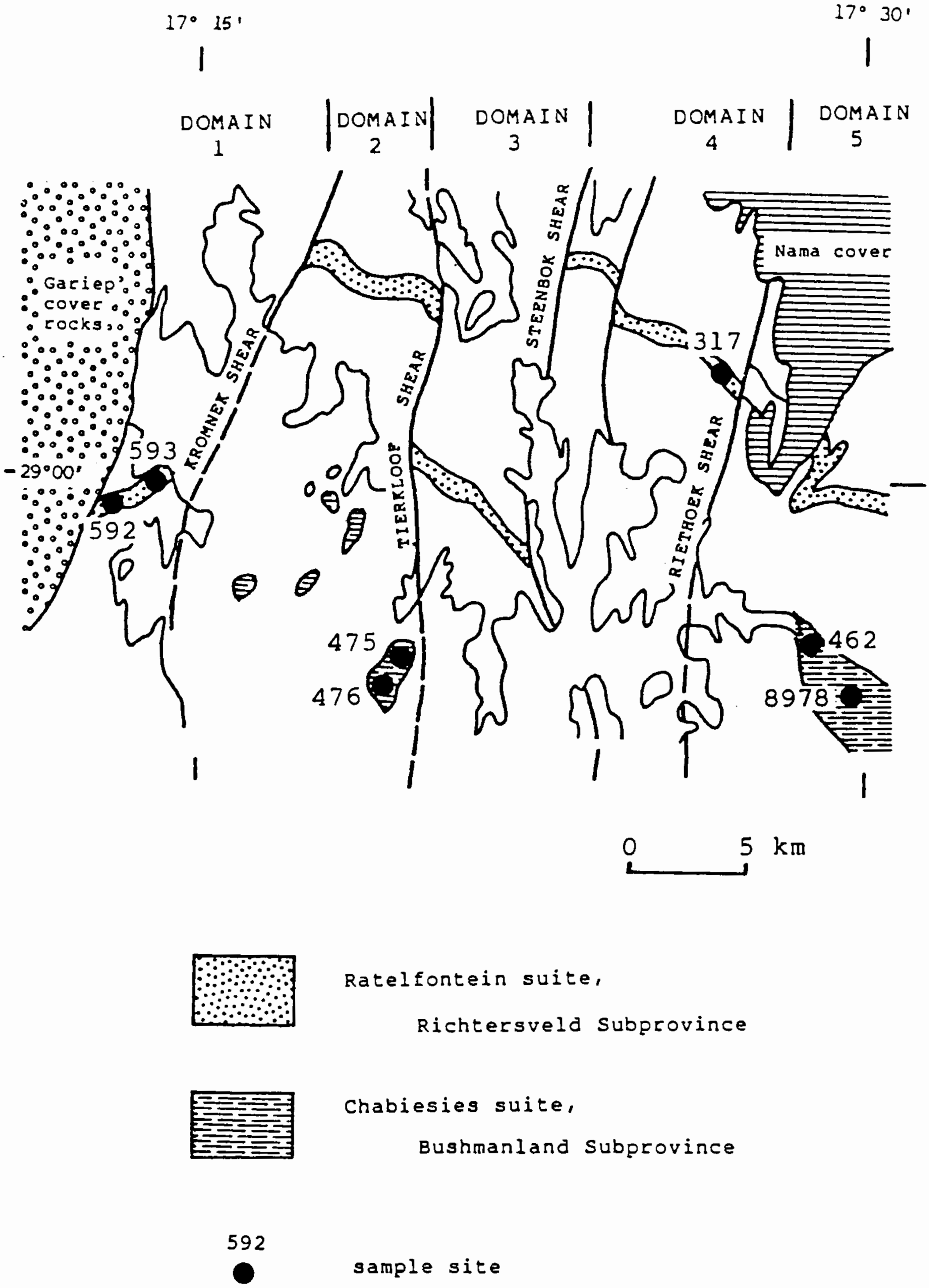


Fig. 5.11 Locality map of garnet-bearing metapelites from the Ratelfontein and Chabiesies suites.

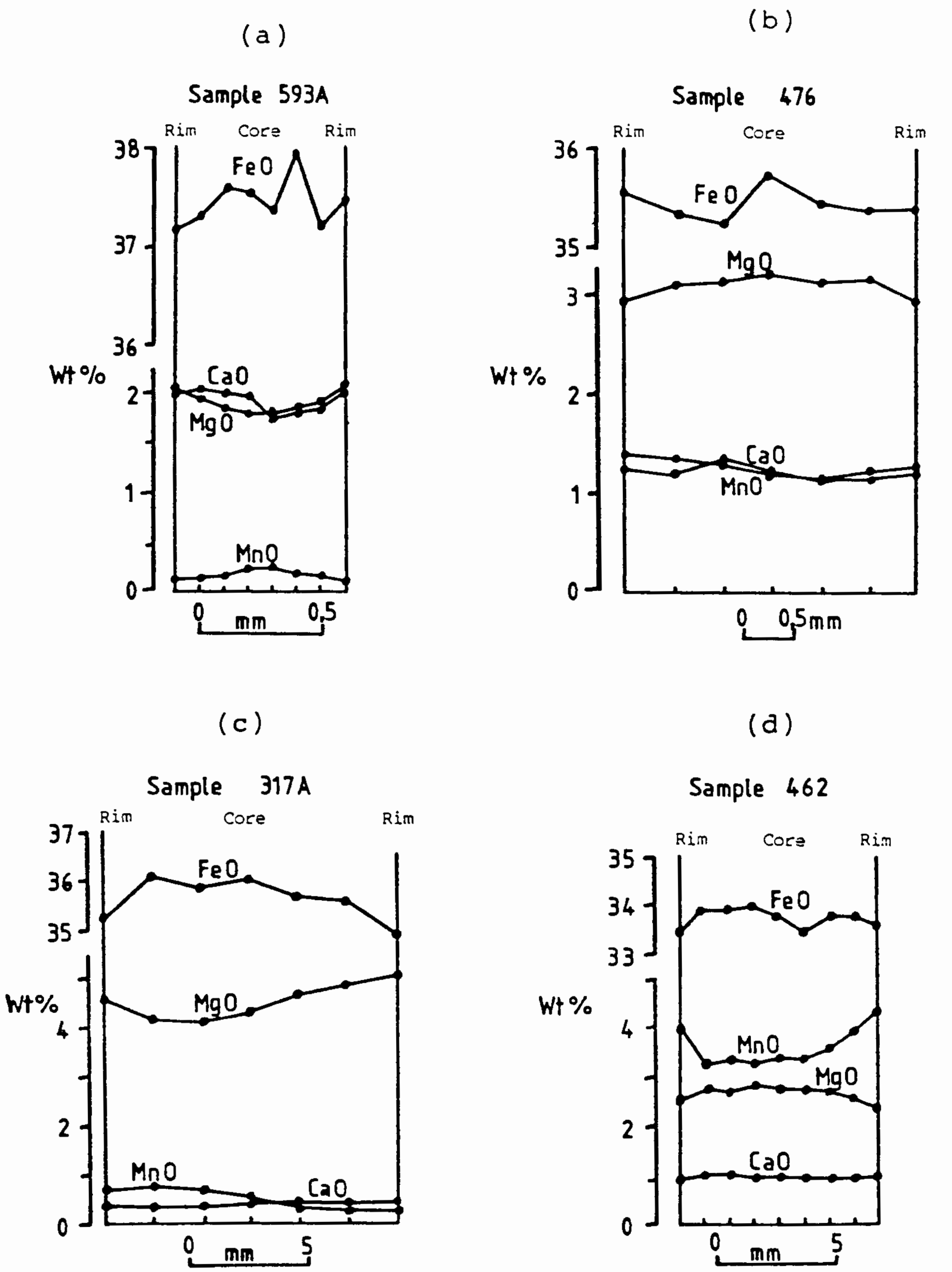


Fig. 5.12 Chemical zonation in garnets.
(a) sample 593A - Ratelfontein suite, Domain 1
(b) 476 - Chabiesies suite, 2
(c) 317A - Ratelfontein suite, 4
(d) 462 - Chabiesies suite, 5

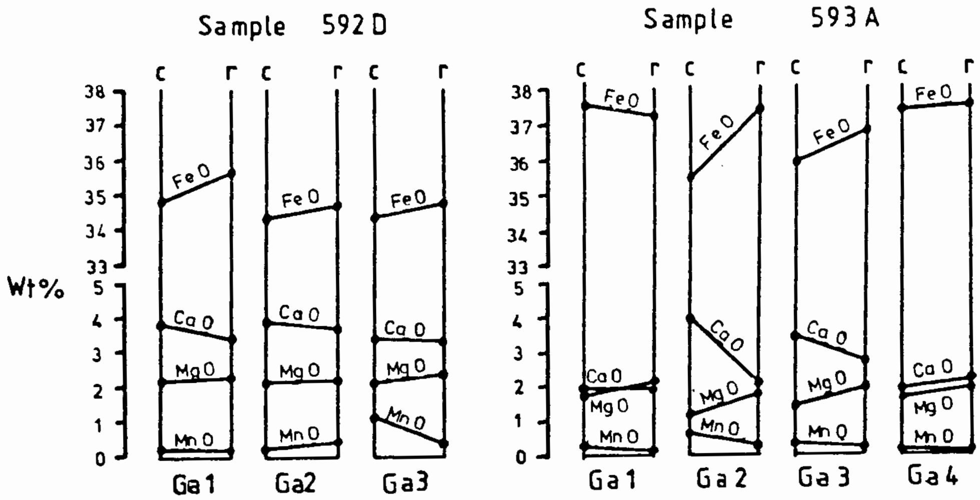


Fig. 5.13 Zonation in garnets, Ratelfontein suite, Domain 1.
C = core, R = rim

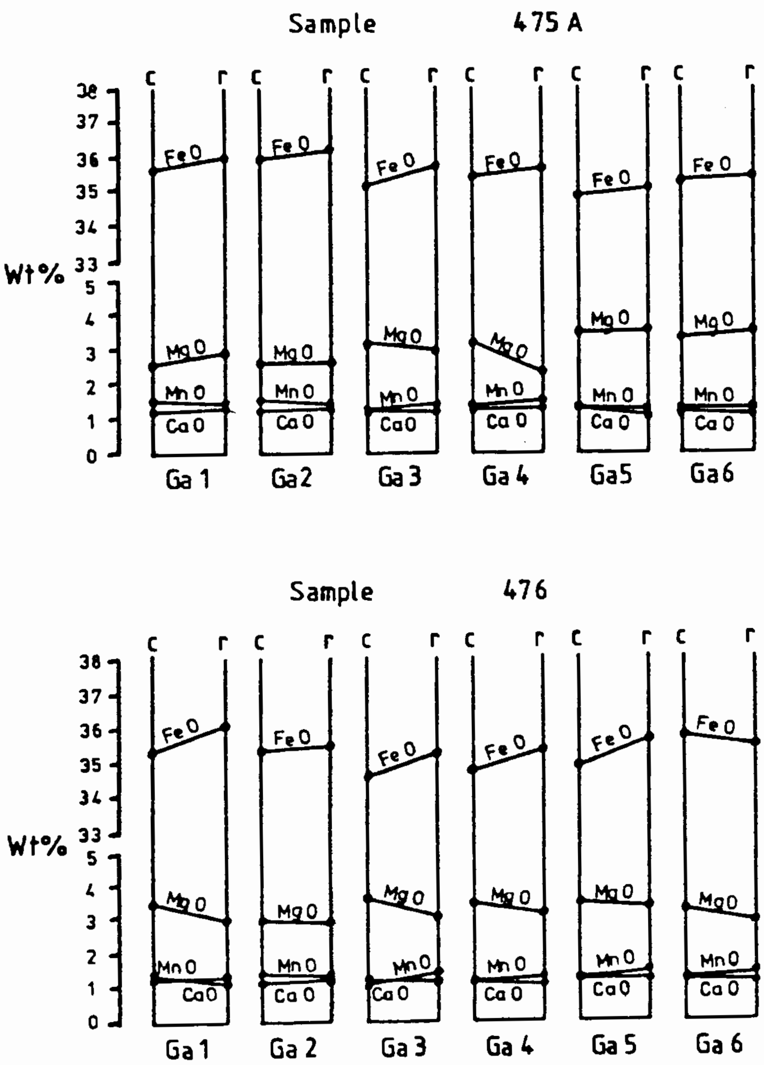


Fig. 5.14 Zonation in garnets, Chabiesies suite, Domain 2.
C = core R = rim

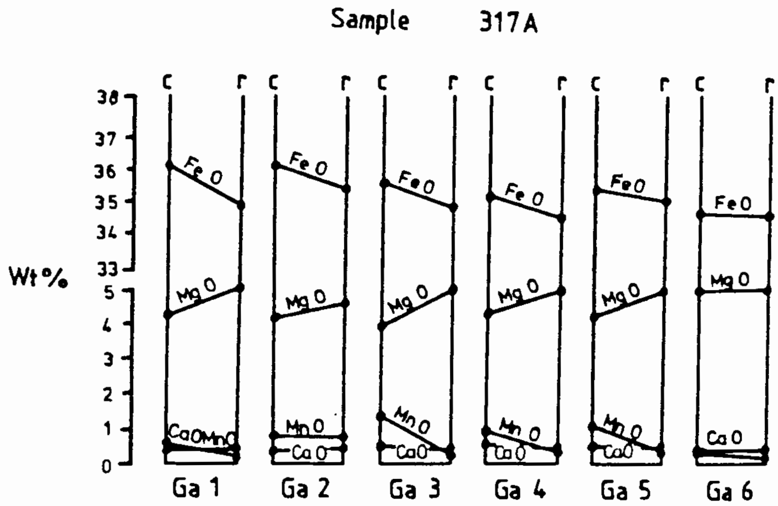


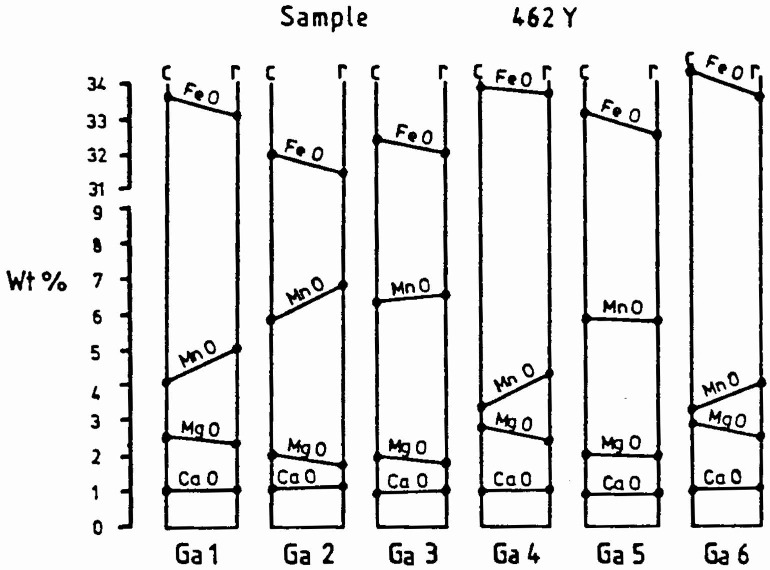
Fig. 5.15 Zonation in garnets, Ratelfontein suite, Domain 4.
C = core R = rim

virtually no variation across the garnet grain. In general profiles in the core regions of the garnets are rather flat, but show a marked deflection in the rims. Garnets from this domain are Mn rich compared to garnets from the other domains (Fig. 5.12).

Interpretations

Garnets from Domain 1 have a simple symmetric zoning showing low Mg, Fe and high Mn in the core relative to the margin. This pattern is typical of "normal" zonation in garnets, and suggests that processes involving depletion and/or diffusion were operative during their growth (Harte and Henley, 1966; Dudley, 1969; Grant and Weiblen, 1971).

The "normal" zoning pattern in these garnets was probably attained during prograde metamorphism, the average Mn content in this case decreasing with increasing temperature. In support of this argument, an increase in MgO and FeO towards the rim of the garnet is similar to that found in "normal" zoning profiles in garnets from other areas (Harte and Henley, 1966; Dudley, 1969). The slight increase in CaO from core to rim may reflect crystallization of garnet under conditions of increasing pressure (Crawford and Mark, 1982).



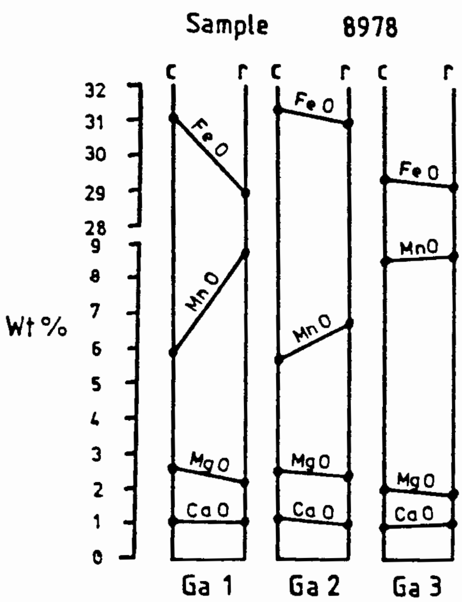


Fig. 5.16 Zonation in garnets, Chabiesies suite, Domain 5.
C = core R = rim

There is little or no evidence for post-crystalline diffusion of Fe, Mn, Mg and Ca in garnets from Domain 1, suggesting that only moderate temperatures (lower amphibolite facies) were reached during prograde metamorphism. This suggestion is supported by petrographic observations of mineral assemblages and textural development in these metapelites (section 5.2.2).

The characteristic bell-shaped Mn profile in garnet is explained by:

(i) Growth during increasing temperature (prograde metamorphism). At low temperatures the larger Mn^{2+} cation is favoured, whereas at higher temperature the smaller Mg^{2+} and Fe^{2+} cations are favoured (Harte and Henley, 1966). The same is true for the Mg/Fe ratio which increases with temperature. An increased proportion of garnet in the rock with increasing temperature and pressure leads to a decrease in the availability of Mn and an increase in Mg content (Miyashiro and Sato, 1973);

(ii) Diffusion involving exchange between initially homogeneous garnet and matrix material in the host rock (Anderson and Buckley, 1973; De Béthune and Laduron, 1975);

(iii) A fractionation-depletion model whereby the strong Mn affinity for garnet is offset by its progressive depletion in the matrix surrounding the growing garnet (Hollister, 1966). This model is based on the refractory behaviour of garnet, i.e. the core of the garnet being isolated from the system during metamorphism means that only the outer rim is in equilibrium with the reactants in the surrounding matrix. The core of garnet therefore formed at different temperatures from that of the margin.

The zoning profiles of Domain 2 garnets show that they are more variable than those of Domain 1. A "normal" zonation pattern is, however, not readily discernable. Garnets in Domain 2 evidently formed at higher temperatures, and probably at greater crustal depth where diffusion processes are likely to be more active, compared to those of Domain 1. This interpretation is supported by mineral assemblages and textural development in metapelites of Domain 2, as well as geothermometry investigations (section 5.2.6).

As in the case of garnets from Domain 1 those from the northern region of Domain 4 show little evidence of post-crystalline diffusion. This suggests that upper amphibolite facies conditions were probably not reached. I propose that changes in both temperature and pressure account for the complex zoning pattern in Domain 4 garnets.

Evidence for prograde metamorphism is based largely on the MnO profile which shows depletion towards the margins, although this is subdued. Diffusion of Fe^{2+} and Mg^{2+} cations in this case had an antithetic relationship during metamorphism, the former showing depletion in the rims whereas the latter shows a steady increase (cf. Figs. 5.12(c), and 5.15).

In my interpretation garnets from metapelites of Domain 4 formed as a result of prograde metamorphism, during the 1200 - 1100 Ma event (section 5.3).

The flattened profiles shown in the core regions of garnets of the Chabiesies suite, Domain 5, indicate that extensive diffusion probably occurred during the peak of metamorphism. MnO may have been concentrated in garnets during this phase. The dramatic increase in MnO towards the rims of garnet grains, accompanied by a decrease in FeO and MgO, is indicative of a "reversed" zoning pattern.

Zoning in garnets is destroyed by internal re-equilibration at higher temperatures, or a levelling off of temperature (Yardley, 1977). At temperatures greater than 600°C stoichiometric and non-stoichiometric material within garnet is redistributed, effectively homogenizing the garnet (Blackburn, 1968; Woodsworth, 1977). "Reversed" zoning in garnet results from processes frozen in during cooling (Grant and Weiblen, 1971; De Béthune and Laduron, 1975; Yardley, 1977; Woodsworth, 1977; Dempster, 1985).

I interpret the "reversed" zoning pattern observed in Chabiesies suite garnets of Domain 5 as having developed during the waning stages of the Namaqua metamorphic event (section 5.3).

Hypotheses put forward to explain the "reversed" zoning pattern in garnet invoke diffusion processes, i.e. during high grade metamorphism garnet no longer acts as a refractory phase but allows cations to penetrate or leave its structure (Grant and Weiblen, 1971; De Béthune and Laduron, 1975; Yardley, 1977; Woodsworth, 1977). Diffusion involves exchange of Mn^{2+}, Fe^{2+}, Mg^{2+}, and Ca^{2+} through the garnet lattice and is a function of grade of metamorphism. The distance over which diffusion is effective in garnet is a function of the diffusion coefficient and duration of metamorphism (Yardley, 1977). An increase in temperature up to 50°C will increase cation penetration rate by a factor of 2 to 4, whereas an increase of 100°C causes the penetration rate to go up by 20 to 40 times (Yardley, 1977). Normal zoning in garnet may therefore be modified at higher temperature where unrestricted diffusion can give rise to a "dezoned" pattern (De Béthune and Laduron, 1975).

Grant and Weiblen (1971) suggest that reversed zoning patterns in garnet are formed either by (i) a drastic increase in Mn into the reservoir during prograde metamorphism, or (ii) resorption of garnet with preferred retention of Mn in the remaining garnet during retrograde metamorphism. In support of (i) above Mn supply could come from the breakdown of ilmenite during prograde metamorphism. Here Mn and Fe are incorporated in the garnet structure, whereas Ti and Fe enter biotite (Hollister, 1969; De Béthune and Laduron, 1975). On the other hand during resorption of garnet, as in case (b) above, some Mn is released into the system, and if not taken up by other phases it then preferentially enters the garnet, diffusing back through the structure. In this way the outer rim of the garnet shows incremental concentrations of Mn, concomitant with lesser Mg and Fe. This results in the characteristic reversed zoning profile, enhanced particularly by Mn (Loomis, 1975). The latter model therefore combines processes of resorption and diffusion (Grant and Weiblen, 1971).

The pronounced reversed zoning profile of garnets in Domain 5 and the antithetic relationship of Fe and Mg towards the rims suggests that homogenization of garnet took place through diffusion processes similar to those described by Grant and Weiblen (1971), and De Béthune and Laduron (1975). This probably reflects the release of Mn during hydrous retrograde resorption of the garnet. Petrographic observations show that the majority of garnet grains in metapelites of Domain 5 have embayed outlines, confirming this interpretation.

The interpretation of the above data from microprobe analyses, in relation to regional metamorphism, can be summarized as follows:

(i) In the east of the study area garnets of the Ratelfontein suite in Domain 4 show characteristics of "normal" zoning, although profiles are quite subdued. I interpret this to be a prograde metamorphic effect during the Namaqua event, where fractional-depletion processes were operative.

In contrast to the pattern in other garnets, those in Domain 5 show profiles of "reversed" zoning. An increase of Mn towards the rims, accompanied by decrease in Mg and Fe is interpreted as the result of limited diffusion near the rim during cooling, after the peak of metamorphism. During homogenization of the garnet the original zoning pattern was probably destroyed (cf. Woodsworth, 1977; Anderson and Olimpio, 1977). It is evident that in this case resorption of garnet took place, Mn cations diffusing into the garnet lattice during a levelling off of temperature. This interpretation is supported by thermometry investigations which suggests that cores of garnets formed at higher temperatures than rims (section 5.2.6);

(ii) In the west of the study area garnets from metapelites of the Ratelfontein suite, Domain 1, show patterns of "normal" zoning. The development of these patterns may be related to the Pan-African event at about 700 Ma, because the Ratelfontein suite here is markedly deformed by shearing along the Stinkfontein Contact Shears.

Garnets of the Chabiesies suite, Domain 2, formed under prograde metamorphic conditions, although subdued zoning patterns probably indicate a complex history of garnet growth.

5.2.3 Metamorphism of basic rocks

5.2.3.1 The Klipbok complex (Domain 4)

Metabasites have the mineral assemblage

muscovite + hornblende + plagioclase + epidote + quartz + clinozoisite.

The mineral assemblage is represented on an ACF diagram as in Fig. 5.17.

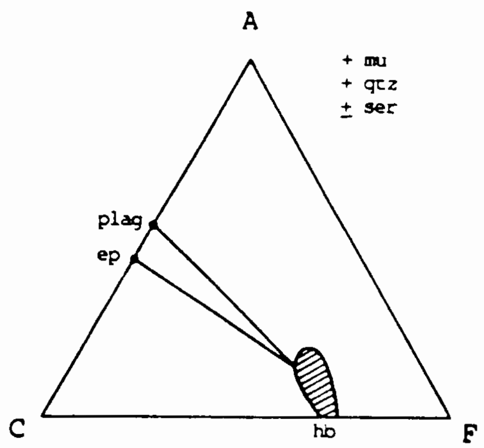


Fig. 5.17 A C F diagram (after Winkler, 1974) representing mineral assemblage in mafic rocks which form in the epidote-amphibolite facies, Domain 4.

Hornblende has a distinctive blue-green colour and a poikiloblastic habit, enclosing small recrystallized quartz grains. Some large hornblende grains have thin needles of ore exsolved parallel to their prismatic crystallographic faces. The other essential mineral is plagioclase (optically determined as labradorite). It usually forms anhedral lath-shaped grains smaller in size than hornblende, and shows varying degrees of replacement by sericite.

Considering the origin of the mafic rocks (Chapter 3) hornblende here has probably formed from original pyroxenes and olivine in the presence of water (Winkler, 1974, p. 163). There is, however, no petrographic evidence to confirm this proposal, as all the original minerals have reacted out to form new minerals, during the Namaqua metamorphic event (1200 - 1100 Ma).

Characteristics of the afore-mentioned diagnostic minerals, viz., the blue-green colour of hornblende, plagioclase with a calcium content > 17%, as well as the presence of clinozoisite indicate that conditions of metamorphism fall within the epidote- amphibolite facies of Miyashiro (1973, p. 305). These characteristics reveal metamorphic conditions during the Namaqua metamorphic event.

5.2.3.2 Other mafic units (Domain 5)

In the south-eastern sector of the area, some 10 to 13 km to the south-east of the Klipbok complex, mafic horizons have the mineral assemblage

muscovite + hornblende + epidote + quartz.

Here the grain size of hornblende is much smaller, and equidimensional, euhedral grain shapes are more prevalent than in those of the Klipbok complex. Hornblende has a blue-green colour, similar to that in mafic rocks of the Klipbok complex, and plagioclase is absent. The mineral characteristics of these essentially hornblende-epidote rich rocks, reveal that metamorphic conditions are similar to those in the northern portion of Domain 4.

A colour change in hornblende from blue-green---> green---> brown indicates a rise in metamorphic grade (Miyashiro, 1973; Binns, 1965). This phenomenon is ascribed to an increase in Ti content and a decrease in ferric iron, and is exemplified in other areas of the Namaqua mobile belt (Blignault, 1977; Jack, 1980; Blignault et al., 1983). In Domains 4 and 5 the colour of hornblende in all mafic rocks does not deviate from blue-green. This indicates that metamorphic conditions in the north-east of the study area reached middle amphibolite facies during the Namaqua metamorphic event. This is supported by petrographic observations of mineral assemblages in meta-pelites in the same area.

5.2.3.3 Gannakouriep Dykes

(i) West of Steenbok Shear - Domains 1, 2 and 3

In Domains 1 and 2 mafic dykes of the Gannakouriep Suite are characteristically very coarse-grained and have the mineral assemblage

biotite + hornblende + plagioclase + epidote.

These rocks have a heteroblastic texture and show complete metamorphic recrystallization. Biotite is brown whereas hornblende is blue-green. The latter mineral has a poikiloblastic habit, the porphyroblasts being randomly oriented and are up to 4 mm in size.

Plagioclase (andesine) grains form short stubby laths between euhedral biotite laths and large, abundant hornblende grains. Minor euhedral apatite is present.

Mafic dykes in Domain 3 have essentially the same mineral assemblage as in Domains 1 and 2, but are characterised by chlorite retrograded on biotite and hornblende. Metamorphic conditions west of Steenbok Shear therefore reached epidote-amphibolite facies after emplacement of the dykes, with retrograde effects noticeable immediately west of the shear.

(ii) East of Steenbok Shear - Domains 4 and 5

Mafic Gannakouriep dykes have the mineral assemblage

biotite + hornblende + plagioclase + epidote.

This assemblage is essentially the same as that in most of the mafic dykes in the area. There are, however, some subtle differences in mineral textures of dykes east of Steenbok Shear compared to those to the west. Dykes in the eastern sector contain green biotite, as opposed to a brown colour west of Steenbok Shear. Hornblende grain-size in dykes in the eastern sector is generally smaller than in the western sector, indicating some degree of recrystallization and growth in the latter area. These subtle differences in mineral textures suggest that metamorphic conditions affecting mafic Gannakouriep dykes in this area, i.e. in the period after ≈878 Ma, were of a slightly higher grade in the western sector. This probably corresponds to the ≈700 Ma event proposed by Allsopp et al., (1979).

5.2.4 Metamorphism of volcanic rocks

(i) West of Tierkloof Shear - Domains 1 and 2 (Pan-African event)

In Domain 1 the following assemblage occurs in metavolcanics of the Windvlakte suite

muscovite + biotite + epidote + plagioclase + microcline + quartz.

Although diagnostic metamorphic minerals such as the amphiboles are absent from this assemblage plagioclase here is oligoclase. In Domain 2 thin units of hornblende gneiss at the base of the Groenrivier suite have the mineral assemblage

hornblende + plagioclase + epidote + iron ore.

This, together with assemblages in metapelitic rocks in Domains 1 and 2, reveals that the metamorphic imprint reached lower amphibolite facies. I interpret this imprint as having formed during the 700 Ma Pan-African event.

(ii) East of Tierkloof Shear, Domain 3 (Namaqua event)

In Domain 3, just west of Steenbok Shear, metavolcanics of the Windvlakte suite north of the Windvlakte Thrust (Fig. 5.1) have the assemblage

albite + actinolite + quartz + muscovite + biotite + epidote.

The presence of actinolite rather than hornblende, in addition to biotite, indicates that these rocks are in upper greenschist facies. Actinolite changes to hornblende at about 500°C in mafic volcanics, corresponding to the appearance of almandine garnet in metapelites at medium and high pressures (Winkler, 1974, p. 166).

South of the Windvlakte Thrust no diagnostic minerals were observed in metavolcanic rocks which could reveal metamorphic grade. There is, however, an increase in grain size towards their southernmost outcrop near the Groothoek Thrust.

The positioning of an isograd marking the transition from greenschist to amphibolite facies in Domain 3 is made difficult because (i) there are no pelitic rocks in the area between the Windvlakte and Groothoek Thrusts, and (ii) pelitic rocks of the Chabiesies suite are offset beyond the southern boundary of the study area, thereby precluding available evidence from critical mineral assemblages.

However, phlogopite-corundum schist immediately south of the Groothoek Thrust is interpreted as forming under lower amphibolite facies conditions, suggesting that the transition from greenschist to amphibolite facies occurs north of the thrust. I propose that the isograd marking this transition is in proximity to the Windvlakte Thrust (Fig 5.1; see section 5.3.2).

(iii) East of Steenbok Shear - Domains 4 and 5 (Namaqua event)

Hornblende gneisses, interpreted as volcanic rocks, occur immediately north and south of the Groothoek Thrust. Those to the north have the assemblage

plagioclase + hornblende + epidote + quartz,

whereas those to the south have the assemblage

plagioclase + hornblende + biotite + epidote + quartz.

The texture of these rocks is characterized by subhedral elongate prisms of hornblende and blebs of epidote in a granoblastic-polygonal groundmass comprising quartz and saussuritized plagioclase. Subhedral hornblende is strongly aligned parallel to $S(N)_2$, having grown along the foliation. Some hornblende gneiss units have a garbenschiefer texture, hornblende conspicuously developing a feathery feature on exposed rock surfaces (cf. Plate 5). According to Spry (1969, p. 269) the garbenschiefer texture is characterized by the presence of poikiloblastic, subidioblastic porphyroblasts of amphibole, in which prisms of amphibole (hornblende) occur as randomly oriented, stellate, radiating aggregates, aligned mostly in the foliation. This "bow-tie" texture develops mimetically along the foliation during post-tectonic crystallization of the amphibole, when stress conditions are no longer prevalent. The size of the hornblende porphyroblasts is attributed to low nucleation rates, and ease of precipitation along the foliation.

A hornblende gneiss unit, different to the others in that it contains large blue-green hornblende porphyroblasts, occurs at the top of the Groenrivier suite (cf. Plate 6), and has the mineral assemblage

hornblende + biotite + quartz + plagioclase + epidote.

I interpret this rock as a product of deformed Vioolsdrif granodiorite which developed during the Groothoek thrusting episode (cf. section 4.5.1.2). Textural features show randomly oriented hornblende porphyroblasts containing numerous small recrystallized quartz inclusions. The majority of the latter are lined up parallel to the elongation of the porphyroblasts although there are deviations to this pattern.

An example depicting different porphyroblast orientations in the same specimen is shown in Fig. 5.18. Within oval-shaped porphyroblasts the alignment of quartz inclusion trails (Si) at an angle of about 30° to the external foliation (Se) defined by biotite, hornblende and epidote (Fig. 5.18(a)), contrasts with that of Fig. 5.18(b) where Si is at an angle of 75° to Se.

In both Figs 5.18(a) and 5.18(b) the foliation is "bowed out" around each porphyroblast. Hornblende porphyroblasts also slightly overgrow Se. I interpret this texture as follows:
(i) The rock initially experienced shearing (cf. Fig. 4.40). This caused some rotation of prior porphyroblasts and a "pushing out" effect of Se around it;
(ii) During prograde metamorphism porphyroblastic growth continued across Se. Thus metamorphism outlasted tectonism.

In the zone of Groothoek thrusting, in Domains 4 and 5, the grade of metamorphism is lower amphibolite facies, as indicated by the presence of blue-green hornblende in metavolcanic rocks.

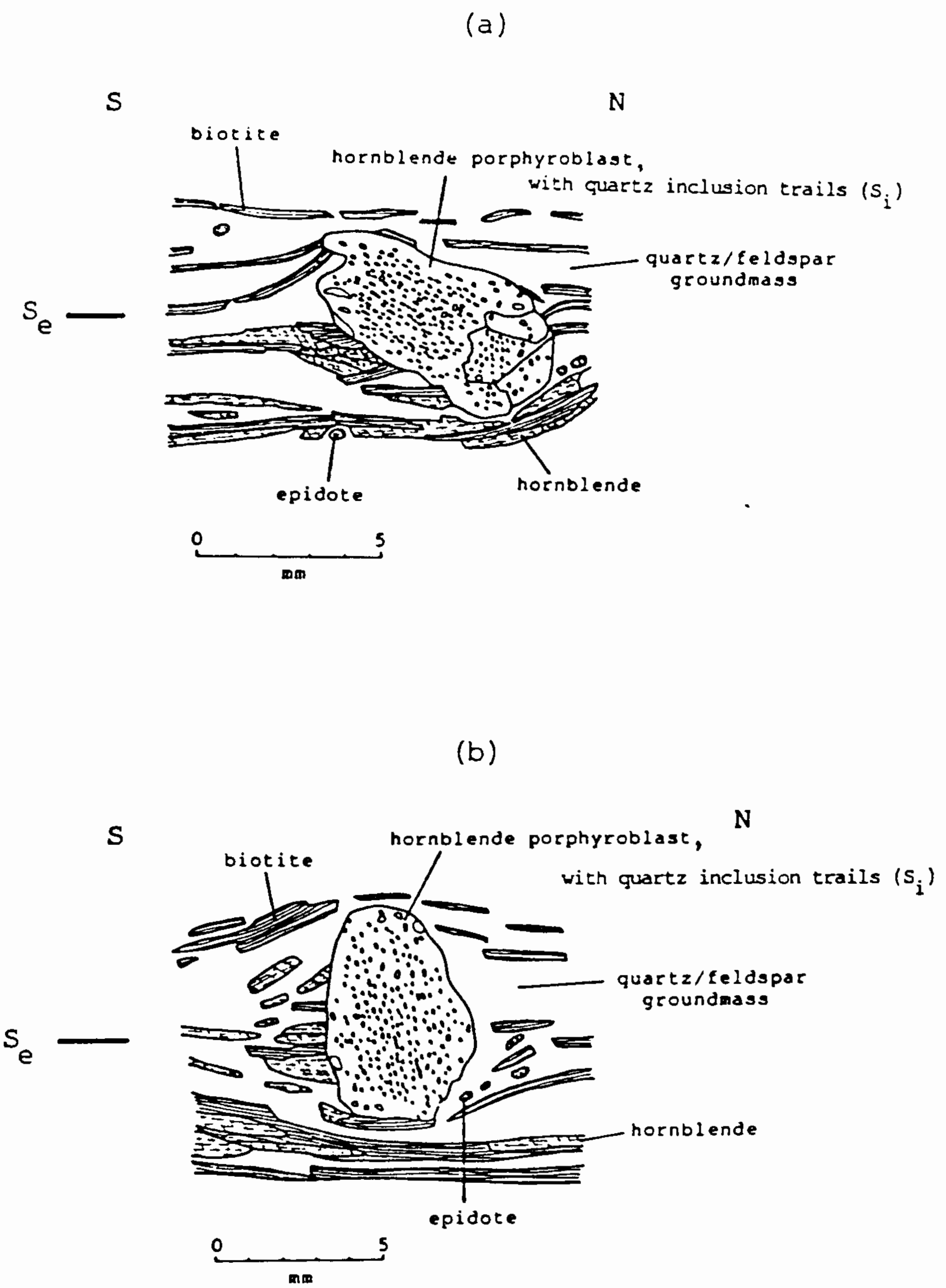


Fig. 5.18 Hornblende porphyroblasts in different orientations in sheared granodiorite, Groothoek Thrust zone.
Inclusion trails of quartz (Si) within porphyroblasts aligned at 30° to Se in (a), and at 75° to Se in (b).

5.2.5 Metamorphism of ultramafic rocks (Namaqua event)

Ultramafic rocks can be metamorphosed to serpentinites by the introduction of H_2O into the system. In this case the original olivine and ortho/ clinopyroxene mineralogy is converted into a serpentinite. However, if CO_2 and H_2O react with an ultramafic rock, then the final product also contains variable amounts of talc + magnesite ± dolomite (Winkler, 1974, p. 148).

Ultramafic rocks of the Klipbok complex and those in the Kouefontein granite gneiss some 6 km to the south are composed of hornblendites and serpentinites. The former are monomineralic rocks, hornblende displaying a characteristic pale green to blue-green colour, and commonly showing a poikiloblastic habit. Euhedral, lath-shaped grains occur in a decussate texture. Serpentinites on the other hand have a greater variety of mineral constituents, being composed of tremolite porphyroblasts (often with exsolved ore lamellae) and pale grey Mg-chlorite, within a ground-mass of serpentine, talc and iron ore.

Serpentine probably formed by the retrograde reaction

$$\text{forsterite} + \text{talc} + H_2O \Longleftrightarrow \text{serpentine} \qquad (8)$$

(Winkler, 1974).

The stability fields of serpentine, forsterite and talc are shown in Fig. 5.19.

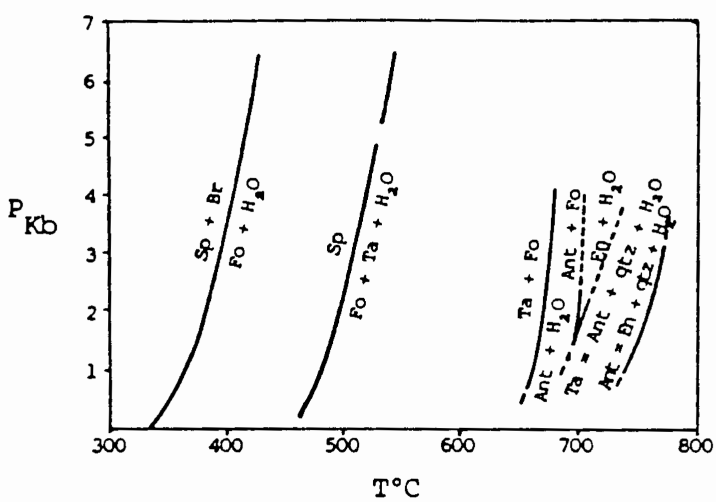


Fig. 5.19 Equilibrium curves showing stability fields of serpentine, and forsterite + talc (after Winkler, 1974, Fig. 11.7).
Sp = serpentine; Br = brucite; Fo = forsterite; Ta = talc; Ant = anthophyllite; En = enstatite; qtz = quartz.

If components such as Al_2O_3 and/, or CaO, are present in the system, this gives rise to new minerals. Mg-chlorite in the Klipbok ultramafics, for example, can be explained by an excess of Al_2O_3 in serpentinites. Chlorite in serpentinites is stable to the high temperature range of medium grade metamorphism, depending on the associated minerals (Winkler, 1974, p. 155).

Tremolite in the Klipbok serpentinites possibly formed by the reaction

$$\text{diopside} + \text{enstatite} + \text{quartz} + H_2O \rightleftharpoons \text{tremolite} \qquad (9)$$

(Boyd, 1959).

There is, however, no petrographic evidence to confirm this suggestion because all the original minerals have reacted out to form new minerals during the Namaqua metamorphic event.

5.2.6 Geothermometry

5.2.6.1 Introduction

Geothermometry calculations in metamorphic petrology are commonly based on exchange reactions involving large entropy and small volume changes. Such reactions are relatively independent of pressure. They are therefore ideal, especially in areas of low to medium grade metamorphism (Tracy et al., 1976; Essene, 1982).

Geothermometer calculations in this study are based on experimental calibrations of Ferry and Spear (1978), and Indares and Martignole (1985). Ferry and Spear's (1978) geothermometer has been shown to be consistent in low and medium-grade metamorphic rocks (Ghent et al., 1979; Essene, 1982; Dempster, 1985). Geothermometry calculations using Indares and Martignole's (1985) formulation are included in this study for comparison.

Exchange thermometers are usually formulated in terms of a distribution coefficient (K_D), which is defined as the ratio of a certain cation pair between coexisting phases. The general formulation relating K_D and temperature at equilibrium is that proposed by Essene (1982). It was derived as follows;

$$\Delta G = 0 = \Delta H - T\Delta S + (P-1)\Delta V + RT\ln K_D + RT\ln K_\gamma$$

where
- G = Gibb's free energy
- H = Enthalpy
- T = Temperature, in °K
- S = Entropy
- P = Pressure, in bars
- V = Volume
- R = Gas constant
- K_γ = Ratio of activity coefficient in the same form as K_D

5.2.6.2 Garnet-biotite exchange thermometer

29 garnet-biotite pairs were analyzed by the electron microprobe for geothermometry determinations (Appendices C-2 and C-3). Rocks containing assemblages suitable for this purpose were found to be impersistent or lacking in key horizons across the study area. Nevertheless, some samples containing appropriate mineralogy were located in critical areas for temperature determinations (cf. Fig. 5.11). No garnet-cordierite determinations could be made as all cordierite grains in samples from the Ratelfontein suite were found to be completely pinitised. Temperature estimates therefore rely totally on ga-bt exchange thermometry techniques applied to metapelitic rocks.

There is a potential problem with Mg and Fe thermometry in using microprobe analyses and assuming all iron to be ferrous iron. Analyses of total iron are recorded by the microprobe as FeO and not Fe_2O_3. Appreciable substitution of Fe^{3+} in biotite yields higher apparent temperatures. In addition there are many possible cation substitutions in garnet and biotite and a unique solution is therefore not possible (Thompson, 1976; Ghent et al., 1979). For example, K_D decreases with increasing Mn and Ca substitution in garnet, the larger Fe^{2+} ion being preferentially incorporated in the garnet structure over Mg^{2+}. This tends to give lower apparent temperatures values (Dallmeyer, 1974; Indares and Martignole, 1985).

In biotite, on the other hand, the substitution of Ti, Al, Fe^{3+} for Mg, Fe^{2+} in the octahedral layer causes a reduction in the dimension of this layer. This reduction is partially compensated by expansion in the tetrahedral layer through increased substitution of Al for Si (Dallmeyer, 1974). The resulting structural mismatch is offset if the larger Fe^{2+} ion is preferentially accommodated over the smaller Mg^{2+} ion and thus K_D is not markedly affected. K_D increases with increasing substitution of Ti, Al and Fe^{3+}. Indares and Martignole (1985) recognize the complicated ga-bt partitioning problem in high grade terranes, and propose a correction scheme which takes into account the effects of Ca and Mn substitution in garnet and Ti and Al^{6+} in biotite.

In general, calculations are based on core and rim analyses of garnet, and matrix biotite. In garnets showing prograde zonation rim determinations give temperatures in equilibrium with coexisting biotite. Garnet rims and matrix biotite will therefore give a good estimate of maximum temperature during prograde metamorphism (Tracy et al., 1976). Those garnets which have undergone retrograde metamorphism core compositions yield maximum temperature values.

It is possible that the Mg/Fe ratio of matrix biotite could increase substantially from its former proportions during metamorphism, thus yielding a lower apparent temperature. Interior garnet and matrix biotite will then give a "false" high temperature estimate (Tracy et al., 1976).

Ferry and Spear (1978) base their formulation on the cation exchange reaction

phlogopite + almandine $\rightleftharpoons$ annite + pyrope (10)

for which the following expression applies;

$$\Delta G = 0 = 12454 - 4.662\ T(^\circ K) + 0.057\ P + 3\ RT\ln K_D$$

where G = Gibb's free energy
K_D = activity product = (Mg/Fe)ga / (Mg/Fe)bt
R = gas constant = 1.98717 cal
P = pressure, in bars
experimental uncertainty = ± 50°C

This thermometer is affected by additional components in solid solution and its application should be restricted for usage with garnets low in Ca and Mg, and biotite low in Ti (Ferry and Spear, 1978).

5.2.6.3 Results

Temperature determinations of ga-bt pairs from garnet-biotite bearing metapelites were carried out using (i) garnet core and matrix biotite, and (ii) garnet rim and matrix biotite, in each rock sample (Appendices D-1 to D-8).

(i) Domain 1

Applying Ferry and Spear's (1978) geothermometer to ga-bt pairs in metapelites of the Ratelfontein suite in Domain 1, temperature calculations range between 473° to 479°C in garnet cores to 510° to 517°C in the rims (Table 5.2). Temperatures are about 50° to 60°C lower in each case applying Indares and Martignole's (1985) formula.

This pattern shows an increase in temperature from core to rim. I interpret this to mean that garnets in metapelites of the Ratelfontein suite in Domain 1 grew progressively during prograde metamorphism.

Equilibrium temperature estimates for Domain 1 are given as 514° ± 50°C.

Table 5.2 Mean temperatures for garnet cores and rims, from metapelites, Domains 1, 2, 4 and 5. (Summary of Appendices D-1 to D-8).

Domain	Formula*	Garnet (mean)	Temperature °C at;		
			5 kbar	6 kbar	7 kbar
1	F & S	core	473	476	479
		rim	510	514	517
1	I & M	core	424	428	431
		rim	449	453	456
2	F & S	core	650	654	658
		rim	626	630	634
2	I & M	core	571	575	579
		rim	550	554	558
			3 kbar	3.5 kbar	4 kbar
4	F & S	core	685	687	689
		rim	753	756	758
4	I & M	core	590	592	594
		rim	645	647	650
5	F & S	core	820	823	825
		rim	775	778	780
5	I & M	core	780	783	785
		rim	754	756	759

*
F & S = calculation after Ferry and Spear (1978), equation 7.
I & M = calculation after Indares and Martignole (1985), equation 19.

Samples probed: garnet = 34
biotite = 29

Probe spots per sample: garnet - core = 2
rim = 2
biotite - 2 in centre

(ii) Domain 2

Results obtained from garnet-biotite pairs of the Chabiesies suite range between 650 to 658°C in cores, whereas in rims they range from 626 to 634°C. Applying Indares and Martignole's (1985) formula, temperature values are calculated as about 80°C lower (Table 5.2).

I interpret this pattern to mean that garnets in metapelites of the Chabiesies suite in Domain 2 developed under conditions of lowering temperature.

Equilibrium temperature estimates for Domain 2 are given as 654° ± 50°C.

(iii) Domain 4

Calculated temperatures range between 685 to 689°C for core determinations and 753 to 758°C for those in the rim, applying Ferry and Spear's (1978) geothermometer to ga-bt pairs from garnet- cordierite bearing metapelites of the Ratelfontein suite (Table 5.2). Applying Indares and Martignole's (1985) formula, calculated temperatures range from 590 to 594°C for the cores and 645 to 650°C for the rims (Table 5.2).

Temperature determinations therefore show an increase from garnet core to rim in metapelites of the Ratelfontein suite. I interpret this to mean that these metapelites developed under conditions of prograde metamorphism.

Average equilibrium temperatures for rocks in the vicinity of the Ratelfontein Thrust (Domain 4),based on garnet rim determinations, are therefore estimated as 756° ± 50°C.

(iv) Domain 5

Temperature determinations on ga-bt pairs from garnet-sillimanite-bearing metapelites of the Chabiesies suite range between 820 to 825°C in garnet cores to 775 to 780°C in rims (applying the Ferry and Spear geothermometer, and cf. Table 5.2). Lower temperature values are obtained using Indares and Martignole's (1985) formula, where core determinations range between 780 to 785°C, and those on rims between 754 to 759°C (Table 5.2).

Temperature determinations on metapelites in the extreme south-east of the study area therefore show that temperature values in cores are greater than those in rims. I propose therefore that the progressive growth of garnet in meta-pelites of the Chabiesies suite was accomplished under conditions of lowering temperature.

Equilibrium temperature estimates for Domain 5 are 778°C ± 50°C.

5.2.6.4 Interpretations

In the western part of the study area there is a 117°C temperature increase from Domain 1 to Domain 2, i.e. from west to east. Temperature estimates for these two domains are, however, not directly comparable because they were determined from two different suites which have a spatial separation of at least 10 km within each domain. These estimates nevertheless suggest a pattern which shows an easterly increase in temperature in the western part of the study area.

The temperature values are compatible with lower amphibolite facies; this is consistent with mineral assemblages in metapelites of Domains 1 and 2.

Temperature estimates for Domain 4 are somewhat higher than the stability range of staurolite which occurs as part of the mineral assemblage in metapelites of the Ratelfontein suite (cf. section 5.2.2.1). These temperatures are, however, acceptable if the following are considered:

(i) The relatively high proportion of zinc in staurolites from Domain 4 has probably stabilized the mineral to higher temperatures (Ashworth, 1975);

(ii) Staurolite in the presence of biotite disappears at temperatures of 675°C at 5.5 kbar (Hoscheck, 1969). However, under water-saturated conditions its experimentally determined stability range can be extended to 680-690°C (Richardson, 1968). Staurolite occurring stably with quartz, cordierite, garnet and sillimanite represents a close approach to termination of this phase in rocks of pelitic derivation (Ashworth, 1975);

(iii) Staurolite in metapelites of Domain 4 always occurs enclosed either within cordierite or garnet. This, in effect, shields the staurolite (i.e. an armoured relic) from further reactions, and the mineral survives "metastably", whilst garnet and cordierite are able to grow at slightly higher temperatures.

Equilibrium temperature estimates for Domain 4 are 126° C higher than those of Domain 2, and only 22° C lower than temperature estimates in Domain 5. Although no samples are available from the Ratelfontein suite in Domains 2 and 5 to define regional temperature variations, there does appear to be a temperature increase from north to south in the eastern part of the study area.

Temperature estimates for the extreme south-eastern part of the study are show that the metamorphic imprint reached upper amphibolite facies. This is consistent with mineral assemblages in meta-pelites of the Chabiesies suite in this part of the study area.

5.2.7 Geobarometry

5.2.7.1 Garnet-plagioclase exchange

Pressure calculations are carried out using the ga-plag-ky-qtz geobarometer (modified after Ghent, 1975; 1976) on Chabiesies suite metapelites from Domain 2, as these are the only ones containing appropriate assemblages for ga-plag geobarometry determinations. The equilibrium expression for ga-plag exchange proposed by Ghent (1976) is based on the reaction

3 anorthite ==== 2 kyanite + grossular + quartz. (11)

As ΔV in this case is large the reaction has a moderate slope in P-T space and is therefore suitable for pressure estimates. The equilibrium expression is given by

$$\Delta G = 0 = 15250 - 39{,}07T + 1{,}582\ (P-1) - RT\ \ln K$$

(modified after Ghent, 1976; in Kasch, 1983)

where G = Gibb's free energy
K = activity product = $(X_{Ca}^{Ga}/X_{Ca}^{Plag})^3$
R = gas constant = 1.98717
T = temperature, in °K
P = pressure, in bars

5.2.7.2 Results

Pressure conditions for Domain 2 are calculated as 5.5 kbar using garnet and plagioclase rim values, and temperatures between 626 and 634°C (Appendix D-9).

No calculations of pressure conditions in Domain 1 were carried out because (i) plagioclase-bearing metapelites were not located in the Ratelfontein suite in the extreme west of the area, and (ii) the Chabiesies suite is displaced to a position south of the study area.

In the region east of Tierkloof Shear pressures are estimated at 3.5 kbar. This estimate is based on the presence of high temperature - low pressure mineral assemblages which occur in metapelites of Domains 4 and 5. Immediately east of the study area the low pressure mineral, andalusite, is present in rocks similar to metapelites of the Ratelfontein suite (Ward, 1977; Blignault et al., 1983). This supports the pressure estimates for the eastern part of the study area.

5.2.7.3 Interpretations

Pressure estimates in the study area suggest that there is a regional pressure variation of about 2 kbar from east to west. This may reflect a regional gradient related to the Namaqua event, or higher pressures in the west may be related to the Pan-African event. I advocate the latter interpretation because (i) CaO patterns in garnets of Domain 1 suggest that garnets in the western area grew under conditions of increasing pressure (cf. section 5.2.2.2), opposite to those in the east (cf. section 5.2.2.1), and (ii) the large dip-slip component calculated for the Pan-African Tierkloof Shear suggests that there is an abrupt pressure gradient from east to west across this shear (cf. section 4.3.2.6).

It appears that lithologies in Domain 2 were metamorphosed at greater crustal depths than those of Domain 1, after the D(P)1 episode. The different P-T inferences for these two crustal blocks can be explained by juxtaposition of Domains 1 and 2 during the Pan-African event. Juxtaposition was effected by left-lateral movement along the Kromnek Shear during the D(P)1 episode (cf. section 4.6.4). Domain 1 was subsequently rotated down to the north, about an axis south of the study area during the D(P)2 episode. This movement positioned the higher temperature, deeper parts of Domain 2 (the Chabiesies suite) at a higher crustal level than Domain 1. Subsequent uplift of Domain 1 along the Kromnek Shear during the D(P)3 and D(P)4 episodes was probably not of any great magnitude, as shown by the lower temperature values in Domain 1 compared to Domain 2.

5.3 INTERPRETATION OF METAMORPHIC EVENTS

5.3.1 General considerations

Textural evidence reveals that prograde and retrograde metamorphism are superimposed in the present area, at different stages during the Proterozoic and Early Palaeozoic. The study area is therefore one characterized by polymetamorphism.

Prograde metamorphism takes place when rocks are changed from their present state by an increase in temperature and pressure and, or change in fluid activity. New mineral assemblages are produced through reactions which progressively expel the fluid phase from the rock, and from hydrous minerals such as the micas and amphiboles. The highest grade of metamorphism is then characterized by anhydrous minerals, e.g. the pyroxenes, and orthoclase.

In attempting to establish metamorphic conditions which are responsible for particular mineral assemblages in a rock, it is customary to rely on the results of experimental data. The final assessment of P-T estimates is at best approximate, in view of inherent experimental difficulties when comparing, for example, experimental time to geologic time. In the laboratory experimental difficulties in duplicating Al_2SiO_5 polymorph triple point results are well known (Winkler, 1974, p. 90; Best, 1982, p. 488). Substantial experimental discrepancies are probably due to composition and structural variation of experimental phases compared to those occurring naturally in rocks (Holdaway, 1971; Greenwood, 1972). At least some experimental deviations of the triple point are due to internal irregularities in sillimanite samples, such as stacking faults and degree of fibrolitization (Salje, 1986). For these reasons P-T estimates vary according to which experimental data are used. Thermodynamic data, however, best compliment a triple point for the Al_2SiO_5 polymorphs at about 3.8 kbar and 500°C (Holdaway, 1971).

Retrograde metamorphism involves forming lower grade minerals at the expense of higher grade minerals if fluids are available, and is usually associated with the waning stages of a metamorphic event. It can also take place if a high grade metamorphic rock fails to survive post-metamorphic cooling (Turner, 1968, p. 94). Retrograde metamorphism involves influx of a fluid phase, usually focussed in shear zones. In this way new chemical species may develop mineral assemblages of lower grade in rocks which had previously attained higher grades of metamorphism. Retrograde metamorphism is facilitated by the introduction of H_2O and CO_2 (Schuiling, 1963).

Complex metamorphic overprinting in the study area makes it imperative to correlate metamorphic events and episodes with those of deformation (Table 5.3). To facilitate understanding of the metamorphic evolution during the Proterozoic Fig. 5.20 shows the tectonic setting at about 1000 Ma, i.e. after the Namaqua metamorphic event, but prior to the Pan-African event.

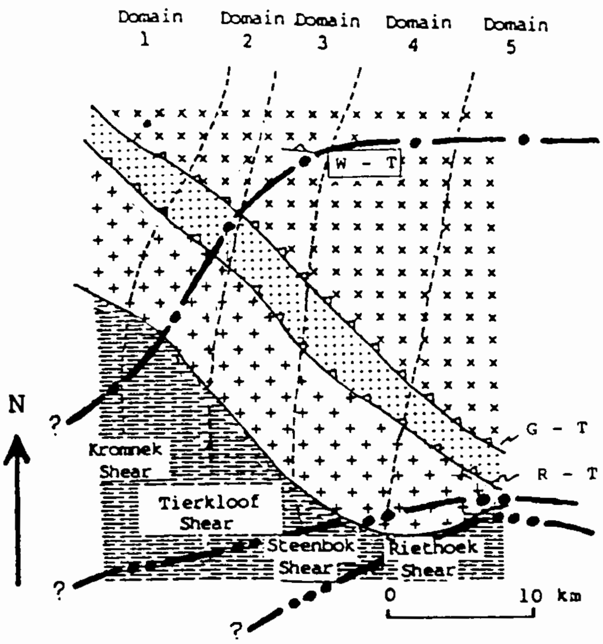
Metamorphic imprints prior to the Namaqua event (1200 - 1100 Ma) are presumed to have been obliterated in the study area because no evidence for such imprints has been found. At least one metamorphic event (M(R)) is proposed to have occurred between 2000 and 1900 Ma, although the grade of metamorphism was probably not higher than greenschist facies (Table 5.3). The Namaqua event is also presumed to have affected the entire study area, although evidence for this is inconclusive in the extreme western part of the area.

The metamorphic imprint during the Pan-African event (at 700 Ma) strongly overprinted that of the Namaqua event in the extreme west of the study area. Mineral assemblages and textural relationships in meta-pelites west of Tierkloof Shear, which differ from those in the eastern part of the study area, are consistent with this interpretation (Chapter 3; and section 5.3.3).

Table 5.3 Correlation of metamorphic and deformational events and episodes in the study area

Era	Deformation event	Deformation episode	Metamorphic event	Metamorphic episode	Metamorphic facies	Proposed time (Ma)
Early Proterozoic	D(R) Eburnian	D(R)1	M(R)?	?	greenschist?	2000 - 1900 (Reid, '79b)
Mid-Proterozoic	D(N) Namaqua	D(N)1				1200 (Clifford et al., '74)
			M(N)	M(N)1	upper greenschist to upper amphib.	1200 -
		D(N)2				1100
		D(N)3				1070
				M(N)2	greenschist	1050
		D(N)4				1000 - 950 (Nicolaysen & Burger, '67)
Late-Proterozoic	D(P) Pan-African: Gariep	D(P)1				750
			M(P)	M(P)1	greenschist to lower amphibolite	700 (Allsopp et al. '79)
		D(P)2				680
				M(P)2	greenschist (in Pan-Afr. shears)	650
Palaeozoic	Pan-African: Post-Nama	D(P)3				530
				M(P)3	greenschist (in Steenbok Shear)	500 (Allsopp et al., '79)
		D(P)4				480
				M(P)4	greenschist (in Steenbok Shear)	470

M(N)1 METAMORPHIC EPISODE - Namaqua event (1200 - 1100 Ma)



Zone index minerals :	Facies of metamorphism :
biotite, actinolite, epidote in metavolcanics	Greenschist
hornblende, andesine in basic rocks hornblende, oligoclase in metavolcanics staurolite, cordierite, anthophyllite, sillimanite in metapelites	lower Amphibolite amphibolite
garnet + sillimanite in metapelites	upper amphibolite

Increase in metamorphic grade

Metavolcanics and plutonic igneous rocks
Ratelfontein, Groothoek and Groenrivier suites } - Richtersveld Subprovince
Sabieboomrante and Kouefontein granites
Chabiesies suite + Gladkop Suite - Bushmanland Subprovince

proposed greenschist/amphibolite transition

proposed ga + st isograd

proposed ga + sil isograd

W - T = Windvlakte Thrust

G - T = Groothoek Thrust

R - T = Ratelfontein Thrust

Fig. 5.20 Interpretation of metamorphic patterns associated with prograde metamorphism during the M(N) metamorphic event. Note future location of northerly-trending Pan-African shear zones.

5.3.2 The M(N) metamorphic event (Namaqua event at 1200 - 1100 Ma)

5.3.2.1 The M(N)1 episode

Evidence for the prograde cycle of the M(N) event is to be found east of Tierkloof Shear, i.e. in Domains 3, 4 and 5. To show the presence of the M(N)1 metamorphic episode, minerals such as chlorite, chloritoid, stilpnomelane and sericite are ignored, and only those forming part of the original assemblages are considered.

The absence of pelitic rock-types north of the Groothoek Thrust, which could confirm diagnostic assemblages, precludes the drawing of an isograd in the northern part of the study area. However, the transition from upper greenschist to amphibolite facies occurs close to the Windvlakte Thrust because the latter is characterized by the greenschist assemblage

biotite + actinolite + albite + epidote

in metavolcanic rocks (northern portion of Domain 3, just west of Steenbok Shear; Fig. 5.20).

The southward transition to amphibolite facies is recorded in Domain 4 by the presence of characteristic blue-green hornblende occurring in the assemblage

hornblende + plagioclase (labradorite) + epidote

in mafic rocks, and

hornblende + plagioclase (andesine) + biotite + epidote + quartz

in intermediate volcanic rocks.

Diagnostic middle amphibolite facies assemblages in metapelites in Domain 4, south of the Groothoek Thrust, include

cordierite + anthophyllite + sillimanite + biotite, and

staurolite + cordierite + garnet + biotite

whereas upper amphibolite facies is interpreted from the assemblage biotite + garnet + sillimanite in metapelites in the southern part of Domain 5 (Fig. 5.20). This is supported by geothermometric determinations (cf. section 5.2.6).

Amphibolite facies is therefore denoted by andesine and hornblende in mafic and intermediate rocks, and staurolite in pelitic rocks, being criteria favoured by Turner (1968). The ga + st isograd is appropriately defined by the southernmost occurrence of staurolite in metapelites east of Tierkloof Shear. This occurs at the base of the Ratelfontein suite, some 15 km south of the Windvlakte Thrust (Fig. 5.20).

The absence of staurolite in Chabiesies suite metapelites in the southernmost part of Domain 5 shows that its upper stability limit has been surpassed; staurolite is therefore unstable in the presence of muscovite + quartz in upper amphibolite facies. The southward increase in grade of metamorphism can therefore be explained by the reaction

$$\text{staurolite} + \text{muscovite} + \text{quartz} \rightleftharpoons \text{garnet} + \text{biotite} + Al_2SiO_5 + H_2O \qquad (7)$$

(Carmichael, 1969; Tracy, Robinson and Thompson, 1976; Kasch, 1987) taking place in these metapelites, under upper amphibolite conditions.

Although lack of sufficient metapelite samples from the south-eastern part of the study area preclude the exact positioning of an isograd, the ga + sil isograd may be located at the top of the Chabiesies suite (Fig. 5.20). I suggest that this position demarcates the change from "normal" to "reversed" zoning in garnet profile patterns which were obtained from microprobe analyses of garnets in metapelites of the Chabiesies suite (cf. section 5.2.2.2). This suggestion is similar to that made by Yardley (1977) with regard to patterns showing changes in garnet profiles in Dalradian Schists, Connemara, Ireland.

The interpretation of isograd patterns shown in Fig. 5.20 is as follows. The M(N)1 episode commenced at 1200 Ma (Clifford et al., 1975) and continued until 1100 Ma. Groothoek thrusting took place over a broad zone of ductile deformation during this episode. Metamorphism, however, outlasted tectonism, as shown by the southward increase in metamorphic grade obliquely across the Groothoek Thrust zone during the M(N)1 episode.

Metamorphic conditions for greenschist (A) and amphibolite (B) facies during the M(N)1 episode are depicted in Fig. 5.21.

Conditions for (B) are based on temperature estimates ranging between 550 and 778°C, and pressures estimated as less than 4 kbar. Andalusite, occurring immediately east of the study area (Ward, 1977; Blignault et al., 1983) confirms that pressure conditions were below 4 kbar, and at crustal depths less than 10 km during the early high-grade metamorphic event. Textural evidence from meta-pelites of the Ratelfontein suite shows that mineral phases during the M(N)1 episode developed during a decompression stage (section 5.2.2.1). I infer that mineral assemblages here formed at peak temperatures even though at reduced pressure, and therefore reflect peak metamorphic conditions (England and Richardson, 1977; Hollister, 1979; Spear et al., 1984).

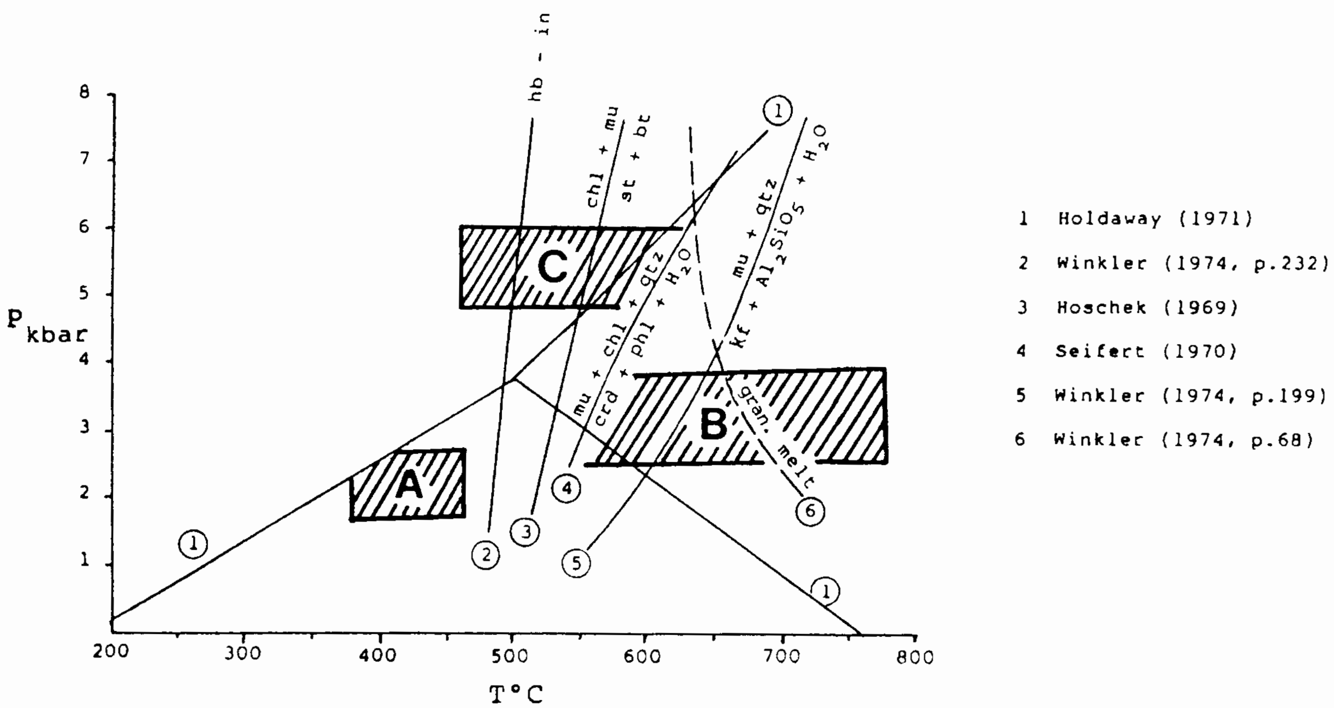


Fig. 5.21 **Phase diagrams relating to M(N) and M(G) metamorphic events in the study area. A = M(N)1 upper greenschist facies, B = M(N)1 mid- to upper amphibolite facies (Namaqua metamorphic event). C = lower amphibolite facies related to the M(P)1 episode (Pan-African).**

These findings therefore confirm Ritter's (1980) original conclusion that there was an early metamorphic event continuous from the Richtersveld Subprovince into the Bushmanland Subprovince. However, the present study suggests that the timing of this event is later than that proposed by Ritter (1980), corresponding with the Namaqua event at 1200 Ma and not an Eburnian-related event (Table 5.3). In addition, Ritter (1980) did not recognize the major tectonic dislocations which have been identified along the marginal zone separating the two subprovinces, during the present study.

5.3.2.2 The M(N)2 episode

A retrograde metamorphic episode M(N)2, which took place after the M(N)1 episode (Table 5.3) is invoked because textural evidence reveals that almost every rock-type in Domains 3, 4 and 5 is variably replaced by sericitization and chloritization (cf., e.g., Figs. 3.3, 3.4, 3.5 and 3.12). This retrograde effect is most pronounced where it coincides with outcrops of the Ratelfontein and lower Groothoek suites east of the Steenbok Shear, in the zone of Ratelfontein thrusting (cf. Fig. 5.1).

Gneisses of the Chabiesies suite in the south east of the study area show complete replacement of broad bands of fibrolite. This occurrence of sillimanite is similar to that described by Grütter (1986, p. 12) in gneisses of the Geselskapbank Formation, some 60 km to the east.

Stilpnomelane probably formed retrogressively in semi-pelites of the Ratelfontein suite in Domain 4, during this episode. In intermediate and mafic rocks of the latter domain textural evidence shows that epidote formed retrogressively from hornblende, sometimes completely pseudomorphing the latter. This is interpreted as occurring during the M(N)2 episode.

Large muscovite porphyroblasts occur in semi-pelitic and pelitic rocks of the Ratelfontein and Groothoek suites and the Kouefontein granite gneiss east of the Steenbok Shear. They are therefore younger than the Kouefontein granite gneiss and probably formed by the reaction

$$\text{K-feldspar} + \text{sillimanite} + H_2O \Longleftrightarrow \text{muscovite} + \text{quartz}. \qquad (14)$$

Alteration of the porphyroblasts along their margins by sericite obscures most of the textural relationships in these rocks. Thus interpretation of muscovite porphyroblast development is tentative. I suggest that the porphyroblasts formed soon after an episode of thrusting associated with the Chabiesies South Thrust (D(N)3; Table 5.3). This episode could well be related to the Skelmfontein thrusting episode in the region to the east (Van der Merwe, 1986).

5.3.3 The M(P) metamorphic event (Pan-African event at 700 - 500 Ma)

5.3.3.1 The M(P)1 episode

There is no relict evidence of the M(N) metamorphic event in the extreme western part of the study area. Mineral assemblages and textural relationships in metapelites of Domains 1 and 2 show that these lithologies developed under different P-T conditions from those in Domains 3, 4 and 5 (Fig. 5.21, shaded area C). The question therefore arises as to why this is the case. Three possibilities can be considered:

(i) The M(N) imprint did not extend across the entire study area;

(ii) Lithologies of the Chabiesies suite west of Tierkloof Shear may in fact represent a sequence younger than 1000 Ma, but older than the Gariep Group (e.g., the Aardvark Group (SACS), some 30 km south of the study area). Metamorphism of this sequence could then be interpreted as a prograde event which occurred after the Namaqua metamorphic event;

(iii) Assemblages which developed during the M(N)1 episode reached only greenschist facies in the western part of the area, but upper amphibolite facies in the south-east of the study area.

Possibility (i) above can be discounted as it is highly probable that the entire area was affected by regional metamorphism during the Namaqua metamorphic event. Possibility (ii) can also be discounted as field mapping shows the stratigraphic succession for both the Ratelfontein and Chabiesies suites is consistent west of Tierkloof Shear. Possibility (iii) therefore appears to be the most likely option. In this case low grade assemblages in the west of the study area which formed during the M(N)1 episode would be overprinted by amphibolite facies metamorphism during the M(P)1 episode. New mineral phases which grew during the medium grade imprint then represent prograde metamorphic products. This occurred in the area west of Tierkloof Shear where lower amphibolite assemblages are present in metapelites at Kabies se Berg, and in sheared rocks of the Ratelfontein suite near its contact with the Stinkfontein Formation. These proposals are supported by an interpretation of the regional pressure pattern, as advocated in section 5.2.7.

Considering the study area as a whole there is a general increase in the grade of metamorphism associated with the M(P)1 episode towards the south-west, based on the following interpretations:

(i) Chloritoid and primary chlorite in the southern part of Domain 5 represent a prograde assemblage (in greenschist facies) grown across broad sericite bands in the Chabiesies South Thrust (cf. Fig. 4.52), and metapelites of the Chabiesies suite (cf. Fig 3.5). In the Chabiesies South Thrust, chloritoid has grown across folded sericite bands, confirming that this mineral phase grew <u>after</u> development and deformation of the characteristic sericite bands in the thrust zone (i.e. after the M(N)2 episode; Table 5.3). Chloritoid and chlorite porphyroblasts in Domain 5 are interpreted as products of metamorphic reactions which may have been brought about by tectonic duplication and burial. I propose that this took place at 700 Ma (Allsopp et al., 1979), after the D(P)1 tectonic episode (Table 5.3);

(ii) West of Tierkloof Shear staurolite and kyanite are zone index minerals characteristic of the lower amphibolite facies. It is possible that these mineral phases formed through prograde reactions (1) and (2), as proposed in section 5.2.2.1. This would require the presence of low grade minerals such as chlorite, chloritoid and muscovite in the basement rocks of this area. These minerals could quite possibly have formed during the M(N)1 metamorphic event, as proposed above, and then reacted out during prograde metamorphism at 700 Ma to form the lower amphibolite facies mineral phases in Domains 1 and 2.

The presence of kyanite in addition to staurolite in metapelites of Domain 2 suggests that prograde reactions similar to (2) and (3), for example, formed these minerals under slightly greater pressure conditions than in rocks in Domain 1.

Fig. 5.22 shows the proposed extent of the metamorphic imprint during the M(P)1 episode in the study area. There are insufficient sample localities to clearly define the M(P)1 staurolite isograd. Staurolites may occur in basement rocks in Domain 3 (i.e., up to Steenbok Shear), but this possibility cannot be substantiated due to the absence of rocks of suitable bulk composition which may yield the crucial information. Such rocks have been displaced along the Steenbok Shear to a position well south of the study area (Joubert, 1971).

At this stage available information suggests that M(P)1 staurolites probably do not occur east of Tierkloof Shear, suggesting that the M(P)1 lower amphibolite facies imprint occurs mostly west of this shear.

On a regional basis I propose that the amphibolite facies imprint in the Hondeklipbaai area (Joubert and Waters, 1980), and farther south in the Sout Rivier area (Waters, Joubert and Moore, 1983), can be extended northwards to include the M(P)1 metamorphic pattern in the western sector of the study area. Farther north M(P)1 grade of metamorphism declines, as shown by the assemblage chloritoid + quartz + muscovite in basement rocks near the Orange River (A.E. Shimron, personal communication, 1986).

I further propose that M(P)1 metamorphism of basement rocks began during the Late Proterozoic, after the D(P)1 deformation episode (Table 5.3). Metamorphic reactions were probably aided by fluids channelled along broad north-east trending shear zones. The influx of water accompanying this penetrative deformation aided recrystallization of schistose rocks west of Tierkloof Shear, and oblit-

M(P)1 METAMORPHIC EPISODE - Pan-African event (700 Ma)

Increase in metamorphic grade ←

Facies of metamorphism:	← lower amphibolite →		← greenschist →
Zone index minerals :	chloritoid staurolite in meta-pelites of Ratelfontein suite hornblende in meta-basites	staurolite kyanite in meta-pelites of Chabiesies suite hornblende in meta-basites	chlorite, chloritoid in meta-pelites of Chabiesies suite

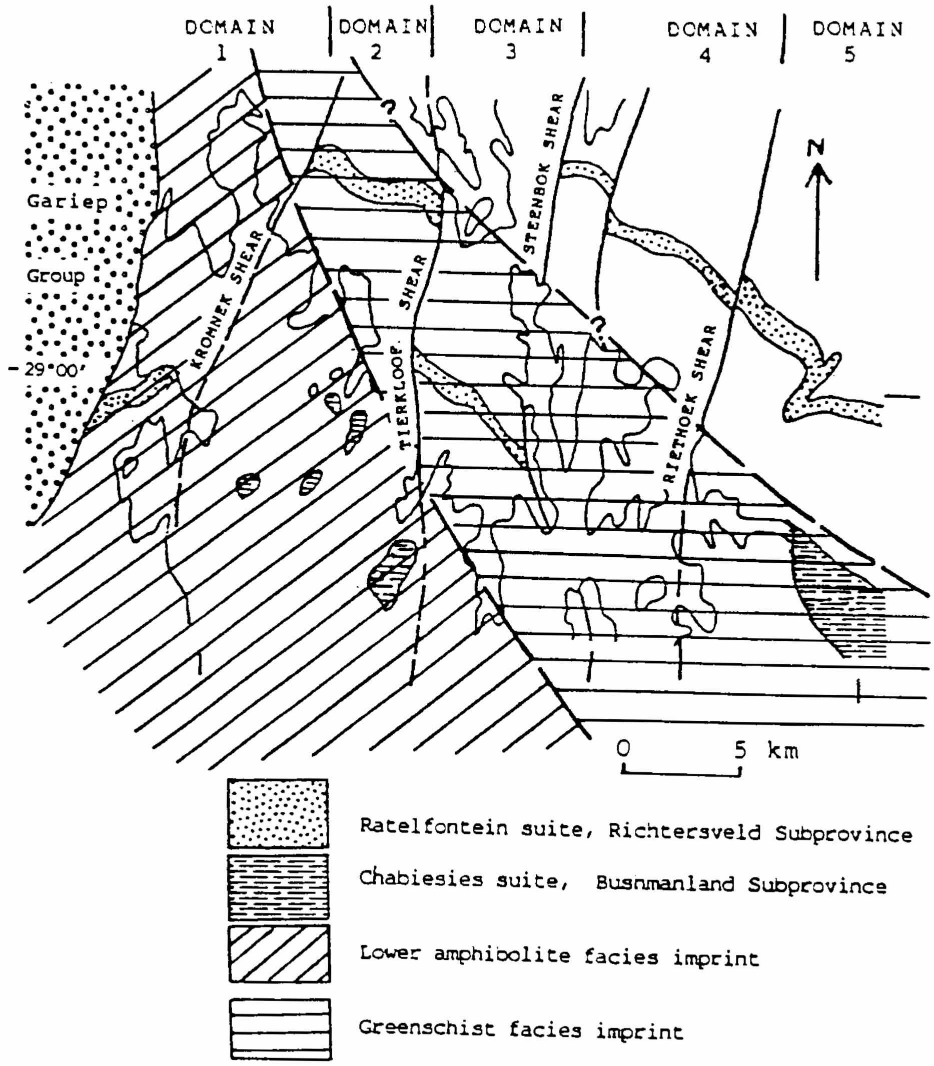


Fig. 5.22 Diagram showing the proposed extent of the M(P) metamorphic imprint in the study area, during the M(P)1 episode. Note that the grade of metamorphism increases towards the south-west.

erated earlier-formed low pressure assemblages. Kyanite-staurolite assemblages in Chabiesies suite meta-pelites west of Tierkloof Shear probably formed at depths of 8-12 km.

5.3.3.2 The M(P)2 episode

This metamorphic episode in greenschist facies is confined to within the Pan-African shear zones which developed during the D(P)2 tectonic episode (Table 5.3).

The rock-type in the majority of the shear zones is a quartz-muscovite schist which has the assemblage

muscovite + quartz ± sericite ± chlorite.

This assemblage is characterized by the recrystallization of quartz into ribbon textures, and muscovite as porphyroblasts and lath-shaped grains. The latter are aligned parallel to the foliation. The porphyroblasts are an earlier generation, their shapes sometimes modified into asymmetric "fish" muscovites during the D(P)2 tectonic event.

5.3.3.3 The M(P)3 episode

This episode is characterised by the development of $S(P)_3$ muscovite cleavages in the Steenbok Shear (cf. Fig. 4.20(a)), associated with the D(P)3 episode at ≈500 Ma, after deposition of Nama sediments (Table 5.3).

This episode took place after metamorphism of the Nama Group. In the south-eastern Richtersveld metamorphism of the Nama Group lies between diagenesis and very low grade (Ritter, 1980). During the present survey no systematic study of the Nama Group was undertaken except to record deformation products of Nama rocks in the Steenbok Shear (phyllonites). The latter reveal deformation under lower greenschist facies metamorphism.

However, Onstott et al. (1986) record a $^{40}Ar/^{39}Ar$ age of 542 ± 4 Ma for a Gannakouriep dyke near Vioolsdrif, which they interpret as a remagnetization age. It is possible that this age might correlate with a metamorphic overprint of similar age recorded by Waters, Joubert and Moore (1983) in the area south-west of Bitterfontein. If this is correct then the northward extension of this younger metamorphic overprint should traverse at least the eastern portion of the present area.

5.3.3.4 The M(P)4 episode

Lower greenschist facies minerals formed during this episode, and are associated virtually entirely with the Steenbok Shear. Chlorite is more prevalent in the Steenbok Shear than the other shear zones and is associated with the D(P)4 episode of uplift along this shear. Chlorite here probably results from hydrolysis reactions during shearing, under greenschist facies metamorphic conditions. The break-down of chlorite into sericite in this shear zone could have produced iron sulphide minerals associated with sulphide gossan occurring at the northern end of the Steenbok Shear. According to Beach (1980), large volumes of H_2O introduced into dilatant shear zones as a result of movement along the shear causes hydration/ hydrolysis reactions resulting in extensive mineralization. Uranium, known to be present along this shear zone, probably precipitated during this phase.

Calcite is an important component associated with retrograde metamorphism in the Steenbok Shear, as well as in agglomeratic dykes adjacent to the Riethoek Shear. It is possible that its presence is due to reactions involving CO_2 and H_2O, e.g.,

$$\text{calciferous amphibole} + CO_2 + H_2O \rightleftharpoons \text{Calcite} + \text{chlorite} + \text{quartz} \qquad (17)$$

(Harte and Graham, 1975), or from reactions involving both plagioclase and potash feldspars in granitic rocks, such as occur in Vioolsdrif granodiorite and Kouefontein granite gneiss.

5.3.4 Summary Statement

The metamorphic pattern in the study area shows the imprint of two metamorphic events, viz.:

(i) The Mid- to Late Proterozoic M(N) event which took place between 1200 and 1050 Ma;

(ii) The M(P) event, associated with Pan-African tectonism between 700 and 470 Ma.

Each of the above metamorphic events comprises at least two episodes (Table 5.3). The area is therefore one displaying complex polymetamorphism.

There is no evidence of a metamorphic imprint prior to the Namaqua metamorphic event, although at least one such event is proposed at 2000 to 1900 Ma.

The M(N)1 episode shows an increase in grade of metamorphism from north to south. Evidence for this pattern is best preserved in metapelites and metabasites occurring east of Tierkloof Shear. The upper greenschist facies boundary coincides approximately with the Windvlakte Thrust in the north. South of this thrust zone mineral assemblages indicate a high temperature-low pressure amphibolite facies grade which increases towards the south. Upper amphibolite facies conditions were reached only in the extreme south-east of the area. The timing of metamorphism during the M(N)1 episode extends from 1200 to 1100 Ma - Groothoek thrusting coinciding with this metamorphic episode. The age of Groothoek thrusting is therefore interpreted as earlier than that proposed by Blignault et al. (1983).

The M(N)2 episode is a retrograde phase on higher grade assemblages formed during the M(N)1 episode. During this phase sericitization and chloritization are associated with a spaced cleavage which is most notable in rocks east of Tierkloof Shear. The zone of most intense alteration corresponds largely to the Ratelfontein suite outcrop, just above the Ratelfontein Thrust.

The imprinted M(P) event is strongest in the extreme west of the study area. It shows a westerly to south-westerly increase in grade of metamorphism, during the 700 Ma (M(P)1) metamorphic episode. East of Steenbok Shear growth of minerals such as chloritoid and chlorite as porphyroblasts across partly retrograded higher grade assemblages in metapelites is interpreted as a lower greenschist imprint related to the M(P)1 episode. West of Tierkloof Shear medium temperature-pressure kyanite-staurolite assemblages formed in metapelites prior to deposition of Nama sediments at 600 Ma. These assemblages are interpreted as a prograde development during the M(P)1 episode. Tectonic loading may have caused burial to greater depths which in turn could have brought about metamorphism (the M(P)1 event).

This metamorphic pattern indicates that the post-Namaqua, pre-Nama amphibolite facies imprint recorded by Joubert and Waters (1981) in the Hondeklipbaai area, and by Waters, Joubert and Moore (1983) in the Sout Rivier area, can be extended northwards to include the present area. The northern limit of the amphibolite facies imprint extends northwards beyond the confines of the present area, probably changing to greenschist facies near the Orange River.

The M(P)2 episode in greenschist facies is confined to Pan-African shear zones and related to the D(P)2 tectonic episode (Table 5.3).

The post-Nama M(P)3 event is manifested virtually exclusively in the Steenbok Shear as a muscovite cleavage development (M(P)3 episode at 500 Ma) and related to the D(P)3 tectonic episode (Table 5.3). The M(P)4 lower greenschist facies episode which developed sericite and chlorite is associated with the last minor tectonic movement along the Steenbok Shear (Table 5.3).

6 TECTONIC AND GEODYNAMIC INTERPRETATIONS

6.1 INTERPRETATION OF THE RATELFONTEIN THRUST AS A MAJOR DISLOCATION ZONE

The Ratelfontein Thrust zone immediately south of the Wortel Belt (Joubert, 1986a) is a significant zone of thrusting (cf. Chapter 4). It is along this zone that active margin rocks (Vioolsdrif Suite/Orange River Group) were juxtaposed against those of continental derivation (Een Riet Subgroup), during an arc accretion event during the Early Proterozoic. This important zone of dislocation is therefore interpreted as a cryptic suture. Other significant regional structures such as the Mike nappe 20 km farther north in the Richtersveld (Ritter, 1980) could also be related to this early tectonism.

In this model 1800 - 1600 Ma supracrustals of the Aggeneys Sequence in the Pella and Steinkopf Terranes are deposited as peripheral basin deposits across the major dislocation zone separating the two subprovinces (Moore, 1986). Field evidence in the Pella area is consistent with this interpretation (Colliston, 1983). Lithologies similar in age to the Aggeneys Sequence are not, however, preserved in the western extremity of the zone separating the Richtersveld from the Bushmanland Subprovince (Vioolsdrif Terrane and western part of the Steinkopf Terrane). Such sequences either never formed in this area, or have subsequently been eroded after several phases of reactivation along the Groothoek Thrust zone.

The Early to Mid- Proterozoic period is marked by the prolific emplacement of granite plutons in the study area (Sabieboomrante adamellite gneiss, Kouefontein granite gneiss, Dabbieputs granite gneiss and a leucogranite, Eyams granite). Field mapping and petrographic observations suggest that tectonism probably took place at intervals of approximately 100 Ma, during this period of time.

The following are important in a regional context:
(i) Deformation and metamorphism of the Gladkop Suite in the Namaqua geotraverse took place at ≈1800 Ma (Van Aswegen, 1983). This event is not recorded in the study area, presumably because of overprinting during the Namaqua metamorphic event;
(ii) Syntectonic deformation is associated with the Kouefontein granite gneiss at about 1650 Ma;
(iii) A proposed ≈1500 Ma event is recorded some 10 km south-east of the study area, based on syntectonic deformation in the Eyams granite (cf. section 4.5.1.2);
(iv) A thrusting event at ≈1200 Ma in the Okiep area some 80 km south east of the study area, is accompanied by emplacement of augen gneisses of the Little Namaqualand Suite (Marais, 1981).

On a regional scale the abovementioned factors favour a convergent plate tectonic model for the northwestern part of the Namaqua Province. Strike-slip faulting may have taken place during the Early Proterozoic tectonic evolution of this part of the Namaqua Province, possibly analogous to the early history of, for example, Cordilleran Mountain belts (Coney et al., 1980). It is, however, impossible to gauge the importance or significance of this type of tectonism in the study area because subsequent overprinting by at least one deformation and metamorphic event have obliterated earlier events.

After Early Proterozoic arc accretion in the north-west Namaqua Province Mid-Proterozoic tectonism was accompanied by sheetlike intrusions of granite along major thrusts. This proposal is in line with the suggested development of major thrusts during continued continental subduction in the southern Himalaya (Dewey, 1977; Toksöz and Bird, 1977). However, in the latter region tectonism has occurred over a very much shorter period of time than that proposed for the Mid-Proterozoic in the north-west Namaqua Province.

6.2 INTERPRETATION OF KLIPBOK COMPLEX ROCKS AS OPHIOLITES

Mafic and ultramafic rocks of the Klipbok complex have several features in common with some deformed Phanerozoic ophiolites.

6.2.1 Rock-type associations

Lithologies of the Klipbok complex comprise chiefly amphibolites and serpentinite; these are possible metamorphic equivalents of gabbroic and harzburgite/ lherzolite/ pyroxenite rock-types commonly found in ophiolites (Conference Participants, 1972; Juteau and Whitechurch, 1979). The Klipbok complex, on the other hand, could be interpreted as a transitional zone of mixed lithologies known to occur in some ophiolites (Moores, 1982).

There are similarities in field relationships and possibly lithotypes between the Klipbok complex and the Border Ranges ultramafic mafic complex Alaska. However, the latter is characterized by the presence of ortho- and clinopyroxene (Burns, 1985) - a factor which cannot be confirmed in the case of the Klipbok complex because of metamorphic overprinting during the Mid-Proterozoic. Burns (1985) interprets the Border Ranges complex as having formed in an island arc environment.

The Klipbok complex is intruded by a leucocratic granite gneiss (Leucogranite) which is a common characteristic of many ophiolites (Dewey and Bird, 1971; Conference participants, 1972; Moores, 1982; Avé Lallemont, 1984). However, there is a very much larger proportion of silicic rocks compared to mafic and ultramafic rocks in the Klipbok complex, a situation more typical of mature island arcs than mid-oceanic ridges (Burns, 1985).

A "complete" ophiolite sequence is generally recognized as a three fold division of rocks, although there are problems in defining an ideal sequence (Haoruo and Wanming, 1979). The sequence consists essentially of ultra-mafic rocks, mafic rocks comprising pillow lavas and sheeted dykes, and a thin veneer of predominantly pelagic sediments. The most characteristic ophiolite lithologies which are missing from the Klipbok complex include pelagic sediments, pillow lavas and sheeted dykes. This suggests that the complex may have formed as part of the plutonic core of an intra-oceanic island arc, similar to the Border Ranges ultramafic/mafic complex in Alaska.

Ophiolites occur as obducted rocks in virtually all Phanerozoic mountain belts (Conference participants, 1972; Coleman, 1971, 1984; Gealey, 1977). During tectonism original ophiolite sequences usually undergo modification, so that in their final setting they are recognised as thinned or dismembered units, where parts of the idealized sequence are missing (Conference participants, 1972).

6.2.2 Geochemistry

Amphibolitic rocks of the Klipbok complex have a chemical composition which is neither consistent with tholeiitic olivine basalts nor calc-alkaline varieties; their original geochemical composition has probably been extensively modified during Namaqua tectonic and metamorphic episodes. A comparison of their trace element content with continental and oceanic basalts (Table 3.26) shows no systematic correlation with either a divergent or convergent tectonic setting (Condie, 1976).

Ophiolites are considered to be products of new oceanic lithosphere, formed either at mid-oceanic ridges (Moores and Jackson, 1974; Church and Coish, 1976), or in a back-arc basin or within island arcs (Dewey and Bird, 1971; Miyashiro, 1975; Leitch, 1984; Coleman, 1984). Controversies regarding the site of formation of ophiolites arise through different interpretations of mafic rock geochemistry. Tholeiitic types form at oceanic ridges, whereas calc-alkali types originate in an island arc/Cordilleran type setting. Some investigators, however, interpret calc-alkali chemical characteristics of ophiolites as products of silica metasomatism of originally mid-oceanic ridge basaltic rocks (Church and Coish, 1976). Field mapping in the study area shows that mafic and ultramafic rocks of the Klipbok complex are emplaced in a volcano/sedimentary sequence - the latter interpreted as fore-arc basin deposits. This relationship is not typical of oceanic crust, but characteristic of deep-seated island arc environments (Burns, 1985).

Interpretations of the origin of the Klipbok complex must await more detailed work because present geochemical results are inconclusive.

6.2.3 A linear outcrop pattern

Lithologies of the Klipbok complex have a strike length of 7 km and a width of 1 km. Their outcrop trace can be extended for at least 15 km eastwards to join up with outcrops of similar rocks along strike described by Ward (1977) and Theart (1980). A continuation of this lineament eastwards joins up with thrusts in the Geselskapbank area (Strydom, 1985) and the Wortel Belt in the Pofadder area (Joubert, 1974c). This lineament may demarcate a possible major geosuture of some 150 km strike length (Chapter 4).

Serpentinites of the Klipbok complex occur mostly as lens-shaped entities in various stratigraphic positions in the Groenrivier and Groothoek suites, but also intimately associated with amphibolitic rocks. This suggests that they may be separate intrusions and, or tectonic lenses. Their mode of occurrence may be analogous to serpentinites of the Great Serpentinite Belt in New South Wales. The latter were emplaced along a fracture within an accretionary prism of an arc complex, above a subducting plate (Leitch, 1979). Serpentinites of the Klipbok complex, on the other hand, could be dismembered bodies of an original ophiolite sequence, their present shape and position being the result of modification during a protracted Groothoek thrusting episode.

Interpretations regarding the origin of the Klipbok complex, on field evidence, remain speculative.

6.2.4 Possible inconsistencies with a geosuture interpretation

There are some inconsistencies to be considered, on a regional scale, if the Klipbok complex is interpreted as representing a geosuture zone. These include the following:

(i) Metasediments and metavolcanics of the Aggeneys Sequence occur north of the Groothoek Thrust in the Pella area (Colliston, 1983; Moore, 1986). Although the stratigraphic/structural setting of supracrustals in the latter area is not fully understood, I support the interpretation that they formed as peripheral basin deposits in a post- collision setting (Moore, 1986);

(ii) Rock types interpreted as possible Vioolsdrif Suite related rocks are recorded well south of the Groothoek Thrust (cf. Holland and Marais, 1983;). Although extremely tectonised in field appearance, these rocks resemble the Vioolsdrif granodiorites from north of the Groothoek Thrust. In the Eksteenfontein area one of these granitoid bodies (Sabieboomrante adamellite gneiss) is similar to Vioolsdrif granodiorite, but differs significantly from the latter in its trace element content. This suggests that the Sabieboomrante adamellite gneiss south of the Groothoek Thrust represents a separate suite which intruded supracrustal sequences at a younger stage than the Vioolsdrif suite. This interpretation is borne out by field evidence in the study area. I suggest that granitoids similar in appearance to the Sabieboomrante adamellite gneiss south of the study area also belong to a suite younger than the Vioolsdrif Suite;

(iii) Ultramafic pods, usually of very small size, occur in the Kouefontein granite gneiss 5 km to the south of the Groothoek Thrust, but also in Vioolsdrif granodiorite just north of the Klipbok complex. Their present position in granitic rocks, and lack of a metamorphic aureole, suggest that they have been incorporated as xenoliths within the latter rocks. This interpretation would account for their irregular distribution and possibly their small size, when compared to the Klipbok complex occurrences.

6.2.5 Comparison of Klipbok complex with other geosuture zones

In the first example a comparison of stratigraphic and geotectonic setting can be made between the Klipbok complex/Groothoek Thrust zone and the Teslin geosuture in the Rockies of the Yukon, Alaska. In the northern Rockies, Erdmer (1985) describes three separate allochthons transported eastwards onto the North American continent during the Mesozoic. The allochthons consist of the following three segments:

(i) Simpson allochthon - a granodiorite/monzonite assemblage (top);
(ii) Anvil allochthon - a mafic/ultramafic assemblage;
(iii) Nisutlin allochthon - a quartz muscovite assemblage (base).

The Simpson allochthon is lithologically similar to the Vioolsdrif Igneous Suite. The Anvil allochthon appears to have similar lithologies to the Klipbok-Wortel type of the North West Cape although the latter are metamorphosed. The Nisutlin allochthon can be equated with the Ratelfontein, Groothoek and Groenrivier suites (Transitional zone) of the Richtersveld Subprovince.

The above-mentioned three allochthons show a complex stacking order, as a result of collision between an island-arc complex and a passive continental margin (Erdmer, 1985). The Anvil allochthon represents oceanic lithospheric material tectonically emplaced during obduction.

Mafic and ultramafic rocks of the Klipbok complex are comparable only in mode of occurrence but not in scale or tectonic setting with those of the Indus tectonic zone of the Ladakh Himalaya (Srikantia and Razdan, 1979) and Yarlung Zangpo ophiolite belt of China (Haoruo and Wanming, 1979). Ophiolites in these areas occur as tectonic slices parallel to the regional trend, in Cretaceous to Miocene sedimentary and volcanic rocks. Their strike extent is greater than 1000 km and they occur over a width of some tens of kilometres. This represents at least an order of magnitude greater than the Klipbok complex. No contact metamorphic aureoles or high pressure metamorphism are associated with mafic/ultramafic rocks of the Ladakh area; this is similar to the setting of the Klipbok complex.

Ophiolites of the Himalaya are interpreted as products of oceanic floor obducted onto a passive margin sequence (Srikantia and Razdan, 1979). The tectonic setting of mafic and ultramafic rocks of the Klipbok complex cannot compare with the Himalayan examples because they represent deformation products formed at vastly different crustal levels. I speculate that mafic and ultramafic rocks of the Klipbok complex may have an analogous setting to the circum-Ungava Line in Canada; the latter represents a deeply eroded suture zone (Dewey and Burke, 1973).

In summary, I suggest that mafic and ultramafic rocks of the Klipbok complex were disrupted and modified during a proposed arc accretion phase at ≈1850 Ma, when they became dismembered and later intruded by the leucogranite. The Klipbok complex may represent dismembered segments of an ophiolite complex.

6.3 PROPOSED GEOTECTONIC EVOLUTION OF THE NORTH-WESTERN NAMAQUA PROVINCE - A SYNTHESIS

6.3.1 Early geotectonic setting

Lavas of the Orange River Group (2000 Ma) and cogenetic Richtersveld Igneous Suite rocks represent a calc-alkaline magmatic province which formed along an active margin similar to a modern island arc or Cordilleran setting (Reid and Erlank, 1979; Reid et al., 1987). As an extension to this model I now propose that the related subduction zone dips towards the north (Table 6.1, and Fig. 6.1).

In this model supracrustal sequences (Ratelfontein, Groothoek and Groenrivier suites) must have formed as fore-arc or intra-arc basin deposits during this period of active subduction (Fig. 6.1). These sequences are overlain by volcanics of the Haib Subgroup which have an interfingering relationship with the supracrustals.

Mafic and ultramafic rocks were emplaced at depth mainly into the volcanic pile. This setting is compatible with their proposed 2000

Table 6.1 Proposed Early to Middle Proterozoic tectonic, magmatic and metamorphic development in the study area, and immediate environs.

Era	Orogenic cycle	Deformation event/episode	Metamorphic event/episode	Richtersveld Subprovince	Bushmanland Subprovince	Proposed time (Ma)
↑	Eburnian	D(R) event		Eruption of Orange River Group lavas, interfingering relationship with supracrustal lithologies. Emplacement of mafic/ ultramafic rocks.	Development of supracrustal sequence in the south, e.g. Chabiesies suite.	2000
		D(R)1 episode		Thrusting in volcanics, e.g. Windvlakte Thrust.		
			M(R) event?	Greenschist facies	?	2000?
				Vioolsdrif Suite		1900 (Reid, '79b).
Early Proterozoic				Arc accretion, development of Ratelfontein thrust sheet.		1850
					Intrusion of Gladkop Suite (granulite facies metamorphism?)	1800
				Proposed accretion of Grünau sequence to northern margin of Vioolsdrif Terrane		1730
				Sabieboomrante adamellite gneiss intruded along southern contact of suture zone.		1700
↓				Southward thrusting, accompanied by intrusion of Kouefontein grantite gneiss between Sabieboomrante adamellite gneiss and Ratelfontein Thrust. Intrusion of Dabbieputs granite gneiss and leucogranite		1650-1600
↑					Syn- to Post-tectonic intrusion of Eyams granite along thrust zone?	1500 (Reid and Barton, '83).
	Kibaran (Namaqua)	D(N) event D(N)1 episode	M(N) event	Deformation and metamorphism of Namaqua Province.		1200 (Clifford et al., '81).
		D(N)2 episode	M(N)1 episode	Greenschist to mid-amphibolite facies.	Mid- to upper amphibolite facies.	1200-1100 (Clifford et al., '74).
Mid-Proterozoic		D(N)3 episode		Groothoek thrusting		≈1100
		D(N)4 episode		Composite (spaced) cleavage development throughout Transitional zone (+ Chabiesies South Thrust)		1070
			M(N)2 episode	Sericite developed along spaced cleavage		1050
				Intrusion of Kromnek-Wyepoort granites		1030
				Main phase of concordant pegmatite intrusion		1000 (Nicolaysen and Burger, '67).
↓		D(N)5 episode		Northward extension phase		950

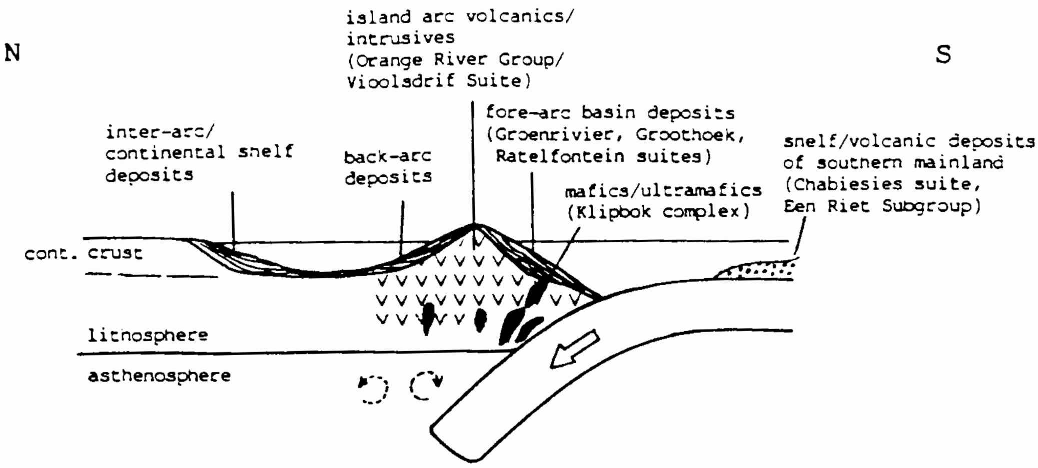


Fig. 6.1 Schematic diagram depicting the geotectonic setting during the period 2000 - 1800 Ma in the Richtersveld region. The main part of the island arc complex is composed of the Orange River Group and Vioolsdrif Suite. The Groenrivier, Groothoek and Ratelfontein suites represent possible fore-arc deposits. Mafic and ultramafic rocks of the Klipbok complex emplaced as arc-related deposits.

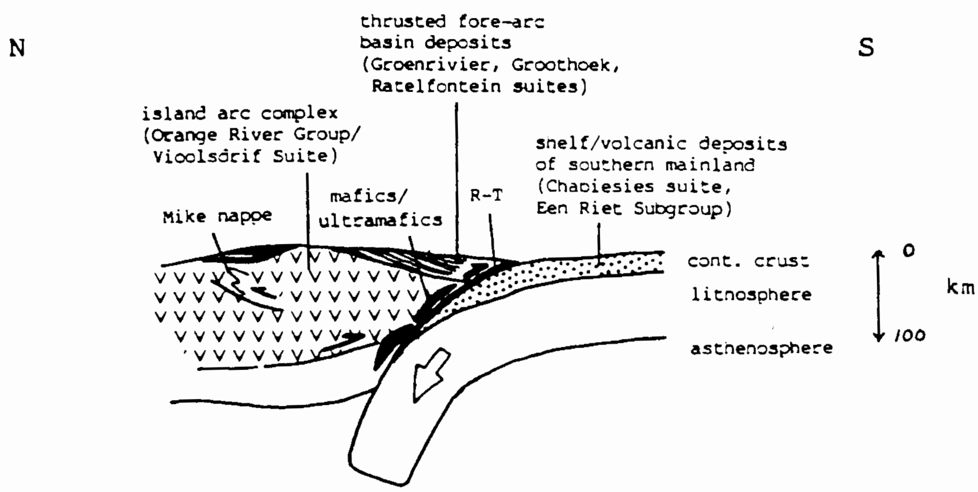


Fig. 6.2 Schematic diagram depicting island arc complex accreted to a southern mainland, at about 1850 Ma. Movement takes place along the Ratelfontein Thrust (R-T), at the base of the island arc complex.

Ma age (Table 6.1). They formed either as primitive ophiolites in a fore-arc region, or small plutons intruded along conduits provided by deep fractures in back-arc regions, during the mature stages of arc development. Geochemical evidence at present is not conclusive with regard to elucidating their initial geotectonic setting. Field evidence, however, strongly suggests that since their initial emplacement into an island arc complex they have become dismembered. They may have formed part of an originally extensive major lineament, the Groothoek "geosuture".

Vioolsdrif granodiorite subsequently intruded volcanic and supracrustal lithologies in the form of large batholiths (Reid, 1979a and b), and as sheets along the southern part of the Richtersveld Subprovince (Chapter 3). Lithologies of the Chabiesies suite (Een Riet Subgroup) are interpreted as a thin passive continental margin sequence.

The first tectonic event (D(R); Table 6.1), occurred prior to intrusion of Vioolsdrif granitoids. Isoclinal folds in volcanic rocks in the northern part of the Vioolsdrif Terrane resulted (Blignault, 1977). Apparently only isolated thrusts (e.g., Windvlakte Thrust) developed in the south during this event (D(R)1 episode).

Earlier metamorphic imprints associated with the D(R) event probably never exceeded greenschist facies.

6.3.2 Arc accretion phase

In the proposed model an island arc complex (Orange River Group/ Vioolsdrif Suite) was accreted to a plate, or microplate (Een Riet Subgroup) at approximately 1850 Ma (Table 6.1; Fig. 6.2). This juxtaposition took place along the Ratelfontein Thrust (Fig. 6.2). A broad imbrication zone formed in the supracrustal sequence of the Orange River Group, above the Ratelfontein Thrust.

Joubert (1986a) proposes a model of crustal accretion where the Groothoek - Wortel line is interpreted as one of the oldest tectonic lineaments in the Namaqua Province (Fig. 6.3). According to the present accretion model this tectonic lineament developed at ≈1850 Ma, and is coincident with the Ratelfontein Thrust, some 4 km south of the Groothoek Thrust.

There is no record of metamorphism associated with this accretion event in the study area, all evidence being obliterated during the Namaqua event (Chapter 5), and possibly during events preceding the Namaqua event. Intrusion and deformation of the Gladkop Suite took place at 1800 Ma (Van Aswegen, 1988), (Table 6.1). It is possible that relict granulites recorded by van Aswegen (1988) in the Namaqua geotraverse south east of the study area formed during

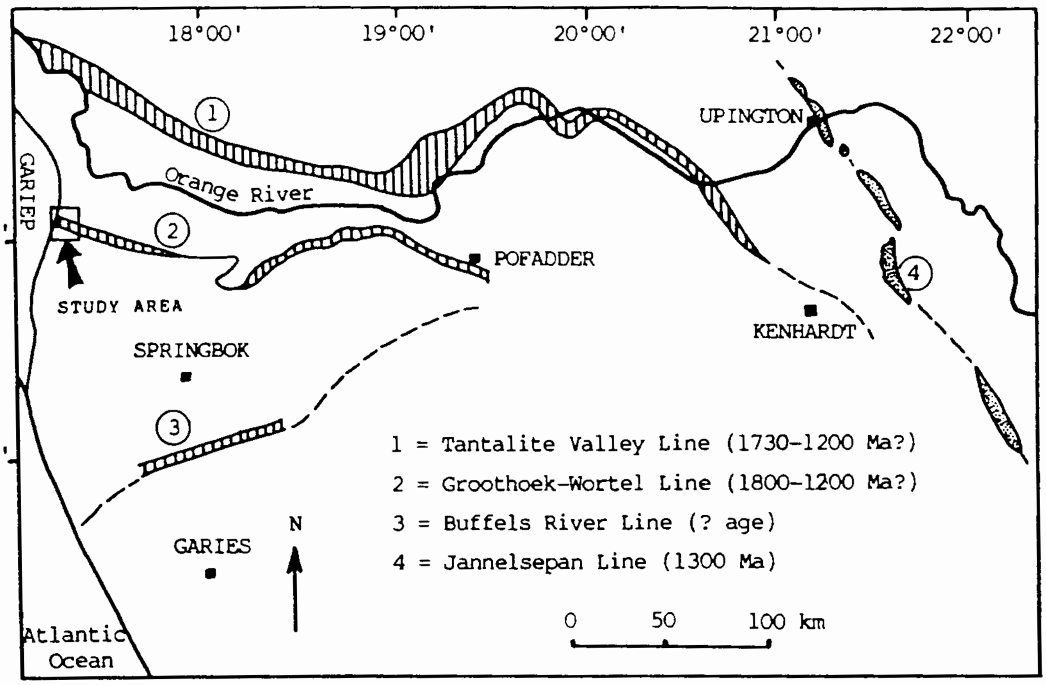


Fig. 6.3 Diagram showing main geosutures in the Namaqua Mobile Belt (simplified after Joubert, 1986a).

Approximate zones of mafic/ ultramafic rocks

Jannelsepan Formation

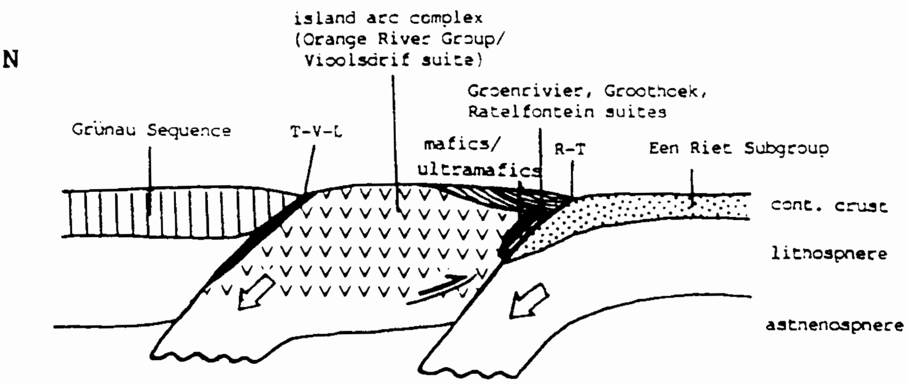


Fig. 6.4 Schematic diagram depicting proposed collision of Grünau Sequence with island arc complex at about 1730 Ma. The Tantalite Valley Line (TVL) marks the suture zone.
R-T = Ratelfontein Thrust.

metamorphism associated with the arc accretion event. During arc accretion subduction probably became modified or ceased because of greater crustal buoyancy brought about by accretion.

I speculate that accretion of the Grunau sequence on to the Vioolsdrif Terrane took place in the north at this stage (≈1730 Ma; Table 6.1), the Tantalite Valley Line marking the suture zone (Figs. 6.3 and 6.4; and Table 6.1). This would agree with the development of the Tantalite Valley Line as proposed by Joubert (1986a).

6.3.3 Magmatic accretion phase

In the study area the Early Proterozoic history, subsequent to 1730 Ma, is characterised by a preponderance of magmatic intrusions. The majority of these intrusions have a sheetlike form and are related in some way to tectonism (Chapters 5 and 6). Magmatic underplating was probably accomplished through modified plate action in an intracratonic setting, as depicted in Fig. 6.5. A mechanism analogous to A-subduction may account for intrusion of the Sabieboomrante adamellite gneiss, although tectonism accompanied intrusion of the granitoid during development of the Sabieboomrante Thrust.

Thrust movement towards the south subsequently took place, accompanied by intrusion of the Kouefontein granite gneiss (Chapters 3 and 4; Table 6.1; Fig. 6.6). The origin of the granite gneiss is probably related to rapid tectonic loading and thickening of crust after accretion of the Grünau Sequence in the north. Thrusts developed into the foreland would produce a stacking wedge and depress lower portions of the crust into an area hot enough for partial melting to take place (Gretener, 1981; Le Fort, 1986). This could induce magmatism which would be assisted by heat input derived from the asthenosphere through convection.

The Dabbieputs granite and the leucogranite probably intruded contemporaneously with or at a slightly later stage than the abovementioned plutons. They nevertheless form part of the magmatic accretion phase.

All of these plutons have a sheet-like form paralleling the southern Richtersveld marginal zone and appear to have intruded along marked structural lineaments analogous to Mesozoic and Caenozoic granitic plutons in the Peruvian Andes (Pitcher and Bussel, 1977). The Sabieboomrante adamellite gneiss may have acquired its tectonic fabric during syntectonic emplacement, in places developing thrusts (e.g., Sabieboomrante Thrust), and has a lens-like form similar to early intrusions in the Andean example.

The origin of the leucogranite which intrudes the Klipbok complex

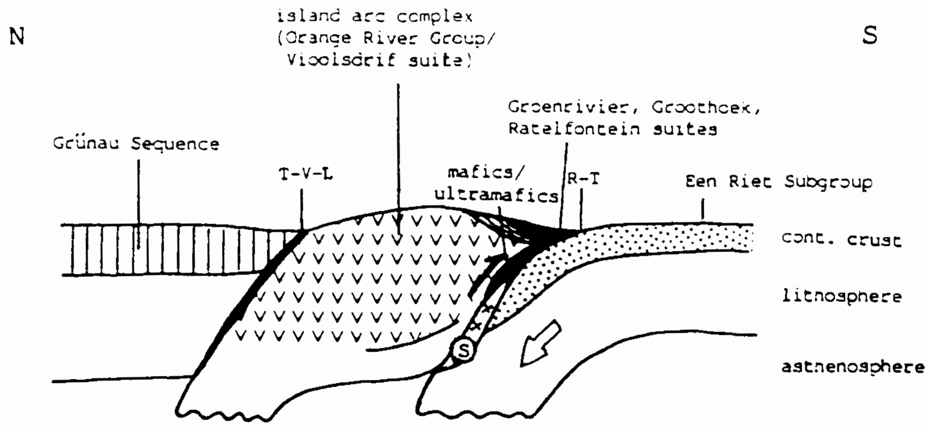


Fig. 6.5 Schematic diagram depicting intrusion of Sabieboomrante adamellite (S) between the island arc complex and the Chabiesies suite, at about 1700 Ma.
TVL = Tantalite Valley Line, R-T = Ratelfontein Thrust.

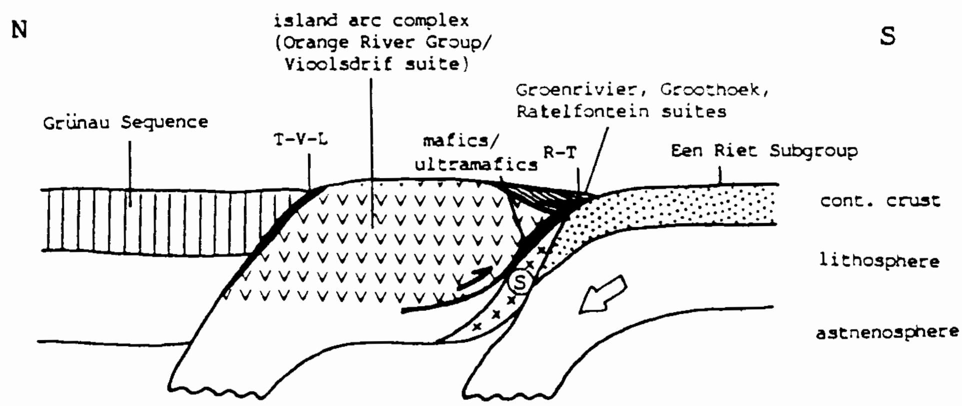


Fig. 6.6 Schematic diagram depicting southward thrusting at the top of the Sabieboomrante adamellite, immediately prior to intrusion of Kouefontein granite between the island arc complex and Sabieboomrante adamellite (S).
TVL = Tantalite Valley Line; R-T = Ratelfontein Thrust.

as concordant sheets remains speculative. A comparison was previously drawn between the structural setting and chemical affinities of this granite with similar leucogranites in the Himalayan mountain belt (cf. section 3.5.5.4). The origin of the leucogranite could be related to a certain amount of shear heating along deep-seated thrusts, and excess metamorphic fluids, similar to that proposed by Searle and Fryer (1986) for leucogranites in the higher Himalaya.

I envisage that the Aggeneys Sequence developed in the Pella and eastern Steinkopf Terranes whilst magmatic accretion was taking place in the southern Richtersveld marginal zone, mostly between 1700 and 1600 Ma.

6.3.4 Development of Eyams Thrust zone

The tectonic history between 1600 - 1200 Ma is not clear. It is possible that after 1600 Ma buoyancy of magmatically and tectonically thickened continental crust prevented further subduction. Subsequent tectonism may have developed new accommodating structures in the form of thrusts in the foreland. It is possible that syn-deformational features in the 1500 Ma Eyams granite some 20 kilometres south-east of the study area may signify such an event, but this proposal remains speculative. Detachment may have taken place along thrusts at metamorphic interfaces, e.g. the granulite/amphibolite interface (Weber, 1986), as depicted in Fig. 6.7. This conforms with the metamorphic grade in the Steinkopf Terrane.

6.3.5 Development of Okiep thrusts

I propose that southward compression took place during the Namaqua event (at ≈1200 Ma), developing thrusts into the foreland in the Okiep region, at greater crustal depths (Fig. 6.8). These thrusts formed contemporaneously with Little Namaqualand Suite augen gneisses (Marais, 1981; and cf. Chapter 2).

Development of the Okiep thrusts probably took place within the zone of granulite facies metamorphism, conforming with the grade of metamorphism in the Okiep Terrane. This proposal is in line with current recognition that thrusts do occur at great depths in the earth's crust (Brewer et al., 1981; Cook et al., 1979; 1981; Oliver, 1982; Smythe et al., 1982; Chadwick et al, 1983).

6.3.6 Development of the Groothoek Thrust zone

The Groothoek Thrust zone formed in response to reactivated tectonism, during the Namaqua event at ≈1100 Ma (D(N)2 episode; Table 6.1). The thrust broke back into the hangingwall, just north of

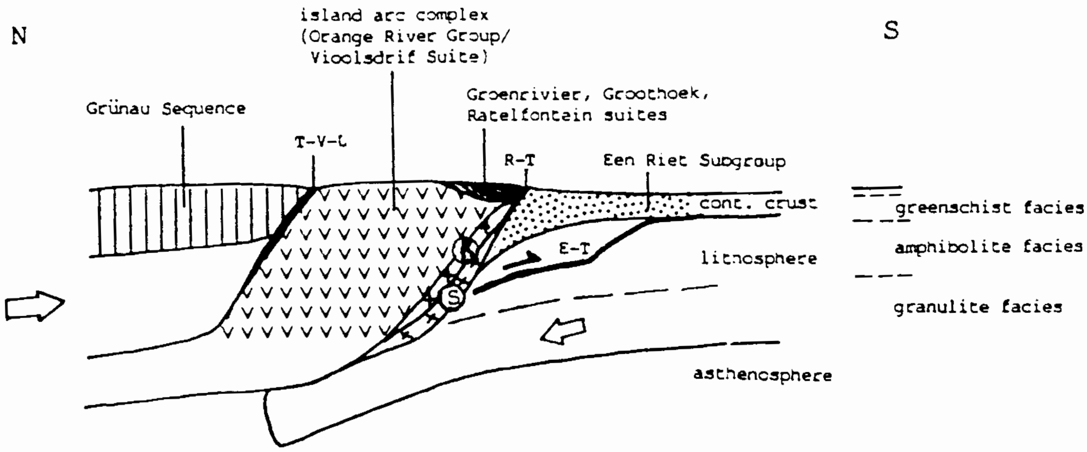


Fig. 6.7 Schematic diagram depicting development of Eyams Thrust zone (E-T), possibly at 1500 Ma. Detachment may have been initiated along the granulite - amphibolite facies interface in the foreland, as depicted in the diagram above.
S = Sabieboomrante adamellite; K = Kouefontein granite; TVL = Tantalite Valley Line; R-T = Ratelfontein Thrust.

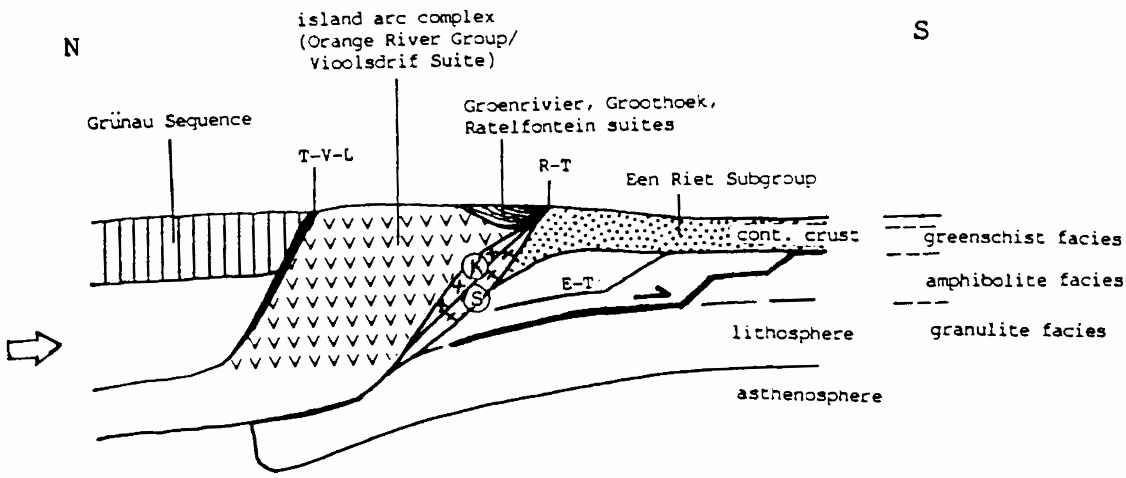


Fig 6.8 Schematic diagram depicting development of proposed Okiep Thrusts (O-T), at 1200 Ma. These thrusts formed in the granulite facies zone, propagating towards the foreland region.
S = Sabieboomrante adamellite; K = Kouefontein granite; TVL = Tantalite Valley Line; R-T = Ratelfontein Thrust; E-T = Eyams Thrust.

the Ratelfontein Thrust, possibly because crustal lithosphere was thickened and downbowed in the foreland region. The thrust formed over a broad zone of ductile deformation, during amphibolite facies metamorphism (Fig. 6.9). Thrusting was accompanied by re-orientation of linear and fold structures, mylonite development, and rotation of hornblende porphyroblasts in hornblende gneiss units. Tectonism was directed towards the south-southwest, indicating a slight anticlockwise swing in the orientation of the stress field during this episode, compared to the Ratelfontein Thrust. This could have been brought about by differential movement along the thrust zone, arising from a build-up of Groothoek Thrust imbricates in the Geselskapbank area (Watkeys, 1988). Such changes in orientation of tectonic stresses with time have been recorded in, for example, the Caledonides (Hossack and Cooper 1986).

In the foreland region a prominent foliation developed in the leucogranite and Kouefontein granite gneiss during this tectonic episode. The Kouefontein mylonite gneiss formed near the base of the latter during this tectonic phase.

The position of the Groothoek Thrust was probably controlled by the following:

(i) The thrust is located at the "dividing line" between massive plutonic rocks and comparably fissile supracrustals. This may be an old suture zone. Tectonic dislocation would favour this interface because old sutures are notable sites of reactivation;

(ii) Crustal thickening of the southern part of the Richtersveld Subprovince initially resulted through imbrication of the Ratelfontein Thrust sheet. During a later tectonic event, e.g., the D(N) event, the base of the competent Vioolsdrif Suite unit would be favoured as a zone of reactivation, because of difficulty in breaching the crustally thickened portion of lithosphere in the foreland.

Decoupling at the "dry" granulite/"wet" amphibolite metamorphic interface at depth would provide a suitable décollement plane along which movement could take place (Weber, 1986). It is also to be expected that ramping up from this detachment plane would be made easier along pre-existing thrust zones or other planes of discontinuity. This implies that reactivation took place at several stages. This is consistent with petrographic evidence. Mafic and ultramafic rocks probably facilitated movement along the Groothoek Thrust during the M(N)1 metamorphic episode (Table 6.1). These rocks acquired their main tectonic features during this reactivation phase.

The first two phases of deformation recorded in the Aggeneys Sequence probably took place during regional overthrusting related to the Groothoek Thrust. Structures such as the Dabenoris nappe

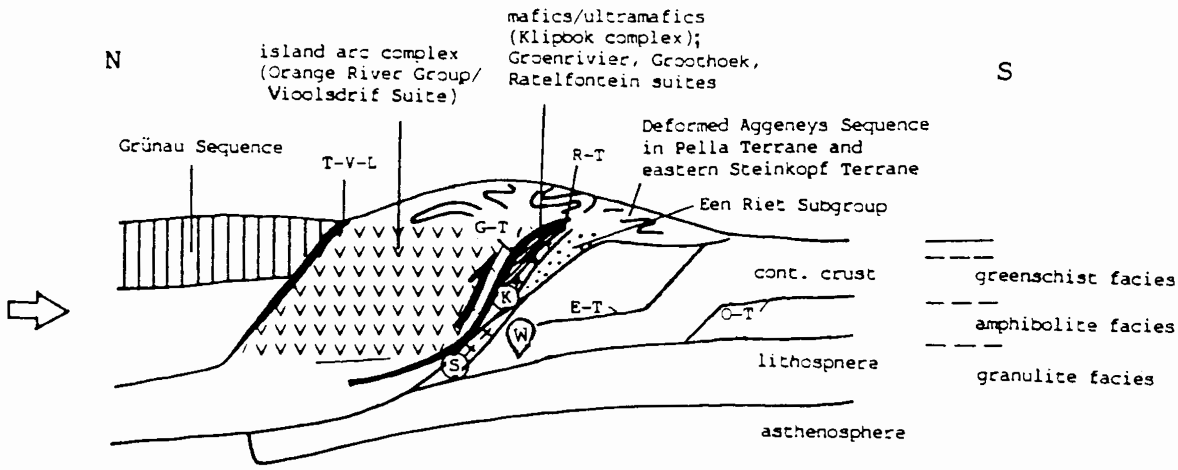


Fig. 6.9 Schematic diagram depicting development of the Groothoek Thrust zone (G-T) at about 1100 Ma. A broad zone of ductile deformation formed at the base of the competent unit of the island arc complex (Vioolsdrif Suite), in response to strong south-southwesterly directed compression.
S = Sabieboomrante adamellite; K = Kouefontein granite; W = K-granites; TVL = Tantalite Valley Line; R-T = Ratelfontein Thrust; E-T = Eyams Thrust; O-T = Okiep Thrusts.

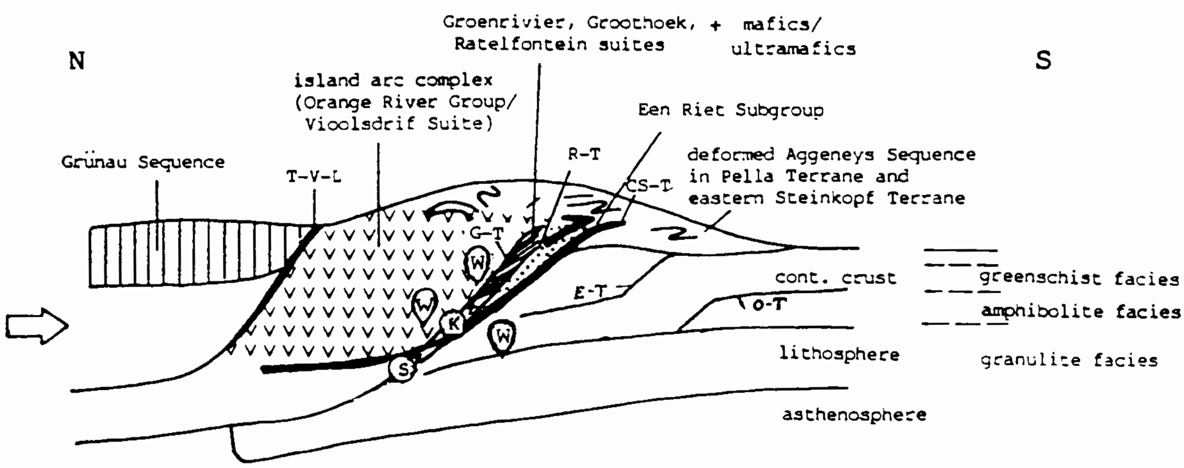


Fig. 6.10 Schematic diagram showing proposed development of the Chabiesies South Thrust (CS-T) at about 1070 Ma. Thrust forms in the foreland, in response to southerly-directed compression.
S = Sabieboomrante adamellite; K = Kouefontein granite;
W = K-granites; TVL = Tantalite Valley Line;
G-T = Groothoek Thrust; R-T = Ratelfontein Thrust;
E-T = Eyams Thrust; O-T = Okiep Thrusts.

formed in the Pella Terrane during this tectonic phase (Blignault et al., 1983).

Timing of Groothoek thrusting is placed at approximately 1100 Ma, as suggested by rotated porphyroblasts in hornblende gneiss units, and modified rock fabric related to the M(N)1 metamorphic episode. This interpretation differs from that of Blignault et al. (1983) in that Groothoek thrusting took place at a slightly earlier stage than that proposed by them.

6.3.7 Development of the Chabiesies South Thrust zone

The Chabiesies South Thrust zone developed in the foreland, in response to further southward compression, during the D(N)3 episode (Fig. 6.10; and Table 6.1). This was accompanied by greenschist facies metamorphism (M(N)2 episode) during the waning stages of the Namaqua event. It seems probable that this episode can be correlated with the phase of post-Groothoek thrusting related to the Taaibosmond/Skelmfontein thrusting in the Namaqua geotraverse (Van der Merwe, 1986; Van der Merwe and Botha, 1989). The Skelmfontein Thrust separates the Steinkopf Terrane (amphibolite facies) from the Okiep Terrane (granulite facies; cf Chapter 2), some 30 kilometres south-east of the study area.

6.3.8 K-granite phase

It is significant that small plutons of K-rich Kromnek/Wyepoort granites are confined to the thickest portion of the crustal segment, and intrude lithologies of the Richtersveld and Bushmanland Subprovinces. Their small size and predominantly rounded form are consistent with the interpretation that they were emplaced during late phases of magmatism, by analogy with plutonism in the Andes (Pitcher, 1978) and other mountain belts (Dewey and Burke, 1973). They may be correlates of the Spektakel Suite (≈1100 Ma), but probably represent a separate intrusive suite (cf. Section 3.2.12).

The K-rich nature of Kromnek/Wyepoort type plutons suggests that they could have formed as differentiated products of lower crustal anatexis, or even as partial melts of mantle material which have migrated upwards through the lithosphere during a post-collision phase. The heat source for their formation could have been provided from the asthenosphere, in the area of greatest crustal depression. Verification of this mode of origin would, however, involve more detailed and specialized studies requiring Sr and Nd isotope data.

6.3.9 Extensional phase

Late Proterozoic northward extension took place approximately parallel to the strike of the Groothoek Thrust zone, but was concentrated along certain zones including outcrops of the Ratelfontein suite, and other planes of weakness. Evidence for this episode is revealed in, for example, micro-structures which show brittle deformation textures in Sabieboomrante adamellite gneiss, boudinaging in the Chabiesies South Thrust, and meso-structures as in chevron folds in the Windvlakte Thrust.

Extension took place after the pegmatite phase at 1000 Ma (Table 6.1). This phase probably corresponds to the same Late Proterozoic extensional tectonic phase proposed by Joubert (1986b) to have taken place to the north of the study area. In the study area, however, the effects of this tectonic event are not manifested to any great extent.

6.4 PAN-AFRICAN EVENT

6.4.1 General aspects

The Pan-African event affected parts of the African continent, including perhaps all of Gondwana, during the period 1000 to 450 Ma (Kröner, 1980; 1982). Kröner (1977b; 1982) suggested that the Pan-African represents a transition phase involving ensialic remobilization, before modern large scale (Wilson cycle) plate movement was initiated.

The region of Pan-African influence pertinent to the present investigation encompasses a north-south strip from Gabon to the Cape, including the Ribeira Belt of Brazil and the north-east extension across Namibia, the Damara Belt. Most authors speculate that the Pan-African event in the south-western African/ South American region began with intracontinental rifting (Martin and Porada, 1977; Barnes and Sawyer, 1980; Miller, 1983; Henry et al., 1988). Two main arms opened up to form the Pan-African South Atlantic (Wilson, 1966; Porada, 1979), but controversy exists as to whether the third arm, the Damara Belt, opened sufficiently to form oceanic crust. Two hypotheses are proposed, viz.,
(i) a tectonic model involving intracontinental rifting and deformation related to the formation of an aulacogen (Martin and Porada, 1977; Porada, 1979; Martin, 1983), and
(ii) a tectonic model involving plate collisions viz. between a northern continental mass (Congo Craton) with a southern one (Kalahari Craton) (Hartnady, 1979; Barnes and Sawyer, 1980; Kasch, 1983). Coward (1983b) proposes an oblique strike-slip model which encompasses aspects of both the above models. Subduction is inferred to have taken place towards the north-west (Weber et al., 1983; Miller, 1983), and possibly towards the north-east (Downing

and Coward, 1981).

Some authors propose that tectonism in the Damara Belt is linked with that affecting the Gariep Group (Davies and Coward, 1982; Kasch, 1983) and the Vanrhynsdorp and Malmesbury regions in the south (Gresse, 1986; Hälbich et al., 1988). An oblique-slip collision model is proposed by the latter authors to account for Pan-African deformation in the Vanrhynsdorp and South-west Cape regions. Most authors emphasize the importance of left-lateral shearing in basement rocks and thrusting in Gariep and Nama cover rocks (Joubert, 1971; Jackson and Zelt, 1984; Gresse, 1986), deformation progressing southwards with time (Hälbich et al., 1988). A composite tectonic model involving strike-slip shearing and thrusting should therefore account for the geological development in all three orogens. There are many examples in the literature where combined thrusting and strike-slip models are invoked to explain the deformation history of orogenic belts (Harris, 1985; Ratschbacher, 1986 ; Gresse, 1986: Ratliff et al., 1988).

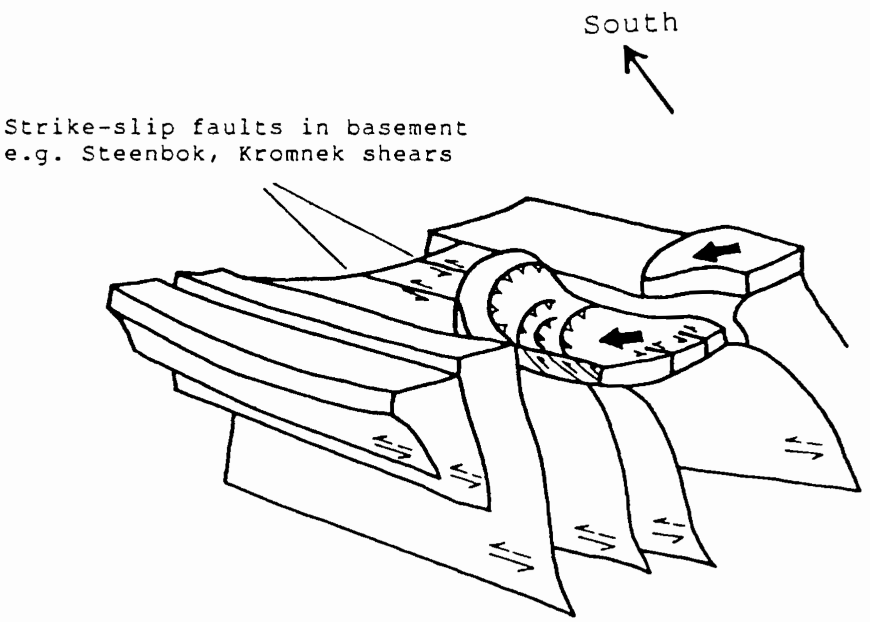
6.4.2 A tectonic model for the Pan-African event south-east of Eksteenfontein

There is an unresolved problem concerning the relation of the 700 Ma age (Allsopp et al., 1979) of the M(P)1 metamorphic episode to early tectonism in this area. Did this metamorphic episode take place during initial Pan-African extension of the western cratonic region, concomitant with sedimentation (Jackson and Zelt, 1984), or during a subsequent tectonic episode? This problem may be solved by invoking a model of combined shearing and thrusting analogous to that proposed by Ratschbacher (1986) to explain the deformation history in the Eastern Alps.

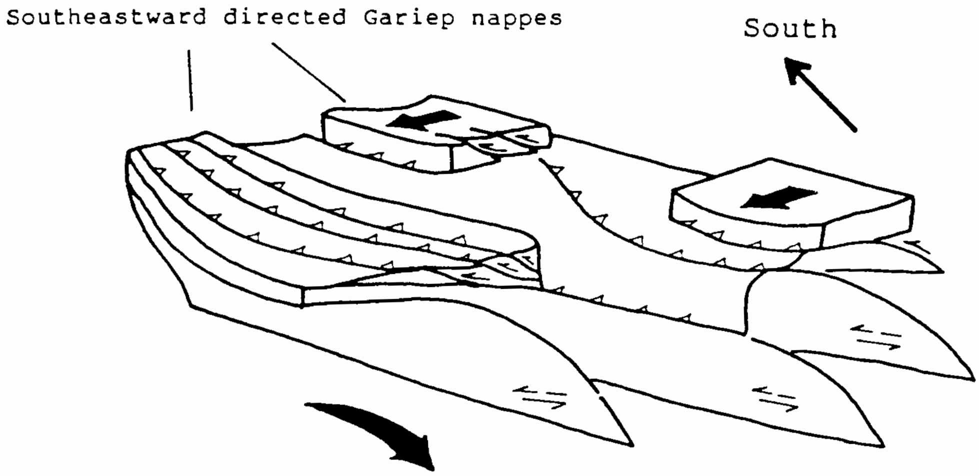
The sequence of Pan-African events as proposed for the study area and immediate environs is summarized in Table 6.2. Left-lateral (transpressional?) shearing in basement rocks began about 750 Ma ago (D(P)1 episode), southward movement taking place mainly along the Kromnek and Steenbok Shears to account for large strike-slip displacements along these shears (cf. sections 4.3.2.7 and 4.6.4). Development of Gariep nappes in the west (Marmora thrust sheet) took place simultaneously with left-lateral shearing in basement rocks in the Eksteenfontein area (Fig. 6.11(a)). Continued movement along these shears during this episode brought about the south-eastward transportation of a pile of thrust nappes at least 8 km thick onto basement rocks in the latter area (Fig. 6.11(b)). This would produce tectonic loading and downbuckling of lithosphere. Such a sequence of events could adequately account for metamorphism of basement rocks in the West Coast Belt (M(P) metamorphic event; Table 6.2). The M(P)1 metamorphic episode was an interkinematic phase which reached lower amphibolite facies in the west of the study area and greenschist facies in the south-

Table 6.2 Pan-African tectonic and metamorphic evolution in the Eksteenfontein area and immediate environs.

Era	Deformation event	Deformation episode	Metamorphic event	Metamorphic episode	Richtersveld/Bushmanland Subprovinces	Gariep Group	Proposed time (Ma)
	D(P)	D(P)1	M(P)		Left-lateral trans-pressional shearing in basement rocks, southerly and south-easterly movement of Gariep nappes.		750
ite oter- oic				M(P)1	Greenschist to lower amphibolite facies	Greenschist facies in cover rocks	700 (Allsopp et al., '79).
		D(P)2			South-easterly directed thrusting, westerly uplift, and rotation of crustal blocks	High- angle thrusting in Marmora sheet	680
				M(P)2	Greenschist facies in north-trending shears	Greenschist facies	650
					Deposition of Nama Group	Erosion of Marmora thrust sheet in the east	600 - 550 (Germs,'83).
rly laeo- ic	D(P)	D(P)3			Left-lateral, trans-tensional movement along Steenbok Shear	Left-lateral trans-tensional movement in Sendelingsdrif-Annisfontein area (Von Veh, pers. comm.,' 87).	530
				M(P)3	Greenschist facies associated with Steenbok Shear	Greenschist facies	500 (Allsopp et al., '79).
		D(P)4			Cleavage development in Steenbok Shear		480
				M(P)4	Chlorite, sericite in Steenbok Shear		470



(a) Thrusts develop in Gariep Group cover rocks during left-lateral, strike-slip shearing in the basement



(b) Continued movement along strike-slip faults in the basement displaces thrust nappes towards the South-East and East.

Fig. 6.11 Schematic diagrams illustrating proposed development of thrust nappes in the Gariep Group (adapted from the transpression model of Ratschbacher, 1986, Fig. 10).
Thrusting is contemporaneous with strike-slip movement in the basement, as depicted in (a). Continued movement along shears transports Gariep cover sequences south-eastward onto basement rocks in the Eksteenfontein area, as in (b). This tectonism initiates metamorphism in basement rocks (M(P)1 metamorphic episode).

east. Nappe thicknesses were greatest in the west, but diminished rapidly east of Steenbok Shear. This is shown by a general increase in metamorphic grade towards the south-west.

The stress field was subsequently oriented towards the south-east, resulting in crustal blocks being upthrust along major shears in the basement, at a proposed time of about 680 Ma (D(P)2 episode; Table 6.2). The southern Richtersveld region probably acted as a pivotal area where large granitic plutons formed buttresses against compressional plate motion. Shears developed in the basement and overlying nappes in a foreland-propagating sequence towards the east, whilst crustal blocks were rotated and upthrust in the west. The Eksteenfontein area therefore behaved as a recess region where fractures steepen in the foreland basement, during their development on the leading edge of a westward subducting plate (Coney, 1973). The transition between greenschist and amphibolite facies at depth (Weber, 1986) possibly formed a suitable detachment plane from which crustal blocks could be uplifted along steep thrusts.

I speculate that fracturing here took place at much the same time as thrusting in the Gariep orogen to the north-west, under greenschist facies conditions (M(P)2 episode; Table 6.2). However, in the latter salient deformation took the form of low-angle thrusting, mainly in cover rocks. Displacement along thrusts in the latter area can therefore be expected to be far greater than in crystalline basement where dip angles on thrusts are generally steep (Burchfiel and Davis, 1975). During the D(P)1 tectonic episode a highland region formed in the west, providing a provenance area for upper Nama sedimentation (Knersvlakte Subgroup; in Gresse, 1986, Fig. 8-4).

Post-Nama deformation took the form of left-lateral, transtensional tectonism, occurring at about 530 Ma (D(P)3 episode; Table 6.2). This tectonic episode affected mainly the Steenbok Shear where a slice of Nama Group sediments was down-dropped and deformed within the shear, during a transtensional movement. Nama sediments were similarly down-dropped along the northern part of the Kouefontein Shear, but do not appear to have been as markedly affected in the remaining eastern part of the study area. West of Steenbok Shear Nama sediments were completely removed after they had been uplifted along major Pan-African shears during this tectonic episode. A phase of greenschist facies metamorphism (M(P)3 episode), emphasized by the development of muscovite cleavages, ensued at about 500 Ma (Table 6.2).

The D(P)3 tectonic episode has similarly been recorded as a left-lateral shear movement in the northern Richtersveld (A.E. Shimron, personal communication, 1986), and a left-lateral transtensional phase in the Sendelingsdrif-Annisfontein area (M.W. von Veh, personal communication, 1987).

An advantage of plate tectonic theory is that tectonic events occurring in one particular area may be related through fault systems to others thousands of kilometres away (Arthaud and Matte, 1977; Tapponier et al., 1986). Applying this principle, and the fact that there are stratigraphic and radiometric age similarities in the Damara, Gariep and Ribeira Belt of Brazil, it is evident that a close correlation and related deformation history exists in all three areas (Porada, 1979; Davies and Coward, 1982; Kasch, 1983; Hawkesworth et al., 1986). Kasch (1983) proposed that oroclinal bending during continental collision caused different subduction rates in the Damara Belt compared to the Gariep Belt. Continental collision in the Gariep Belt consequently occurred slightly later than in the intercontinental branch of the Damara. It is understandable then that tectonic and metamorphic events in, for example, e.g. the Damara Belt need not be exactly contemporaneous with events in other areas.

The earliest Pan-African tectonic episode (D(P)1) recorded in the Eksteenfontein area may be related speculatively to movement on the Purros lineament along the northern Namibian coast. The latter could quite conceivably be linked to faults in the Eksteenfontein area, as part of a larger transform fault system. This proposed fault system extends southward to the Vanrhynsdorp area and to the south-west Cape (Gresse, 1986; Halbich et al., 1988).

The first episode of upthrust-rotational movement (D(P)2) took place at about 680 Ma, and correlates with high-angle thrusting in basement rocks of the northern Richtersveld and within the Marmora sheet (M.W. von Veh, personal communication, 1987). The D(P)3 episode of trans-tensional movement in the recess area at Eksteenfontein (at ≈530 Ma) may well be contemporaneous with deformation in the Naukluft nappes (Weber et al., 1983).

The last tectonic episode (D(P)4) was a minor one, recorded mainly as a cleavage and kink fold development in the Steenbok Shear at about 480 Ma (Table 6.2). This episode may coincide with the last thrusting phase in the Vanrhynsdorp orogen at 480 Ma (Gresse, 1986), deformation being much more pronounced in the latter area.

The last metamorphic episode (M(P)4) took place under lower greenschist facies conditions, with the prominent development of sericite and chlorite, mainly in the Steenbok Shear.

7 SUMMARY AND CONCLUSIONS

7.1 STRATIGRAPHIC RELATIONSHIPS

Supracrustal lithologies have been grouped into suites, equated to stratigraphic formations in rank, because severe deformation and disruption by granitic sheets makes it difficult to map them as a lithological sequence. These are the Ratelfontein, Groothoek, Groenrivier and Windvlakte suites of the Orange River Group (Richtersveld Subprovince), and the Chabiesies suite of the Een Riet Subgroup (Bushmanland Subprovince). Supracrustals of the Richtersveld Subprovince, composed chiefly of quartzo-feldspathic gneisses and minor Mg-rich meta-pelites are diverse, imbricated, and generally lack continuity along strike. They are interpreted as island arc-related deposits (probably fore-arc basin deposits) because of their conformable and interfingering relationships with the island arc-like rocks of the Orange River Group and Vioolsdrif Suite. Supracrustals of the Bushmanland Subprovince (Chabiesies suite) occur in the south of the study area and differ from supracrustals of the Richtersveld Subprovince in two ways; first, their lithological strike is oblique to those in the north, indicating that they have a disconformable relationship with the northern supracrustals; secondly, they do not have a great range of lithotypes. The Chabiesies suite is interpreted as part of a passive continental margin sequence and is therefore not related to those of the Richtersveld Subprovince.

Thick granite sheets (the Sabieboomrante adamellite gneiss and Kouefontein granite gneiss) subsequently intruded along the zone of disconformity which separates supracrustals of the two subprovinces. The boundary between the Richtersveld and Bushmanland Subprovinces is therefore interpreted as a tectonic-igneous one, characterized by strong imbrication in supracrustal lithologies of the former subprovince, and subsequent intrusion of plutonic sheets along the zone of disconformity. This conclusion differs from that of Ritter (1980) who regarded lithogies of the Richtersveld Subprovince as conformable with those of the Bushmanland Subprovince, having undergone the same tectonothermal history.

Volumetrically, intrusive rocks form the greater proportion of rock types in the study area. They include the 1900 Ma Vioolsdrif Granodiorite, a related megacrystic granite and Tweeriviere granite. These plutons are all confined to the northern part of the study area; they form part of the Vioolsdrif Suite intrusive into supracrustals of the Richtersveld Subprovince. The Sabieboomrante and Kouefontein granite gneiss sheets which occur parallel to the north-westerly lithological strike and south of the Vioolsdrif Suite are interpreted as intrusives slightly younger in age than Vioolsdrif Granodiorite. They transect lithologies of the 1800 Ma Gladkop Suite, and occur along a significant structural disconformity separating supracrustals of the Richtersveld and Bushmanland Subprovinces. The Sabieboomrante adamellite gneiss differs significantly from Vioolsdrif Granodiorite in having higher Zr, Y, and Nb, but lower Sr; this suggests that it has formed by crustal contamination and therefore has a different origin to the mantle-derived Vioolsdrif Granodiorite (Reid, 1979a). Even though present geochemical data are inadequate there is enough evidence to propose a different interpretation from that suggested by De Villiers and Söhnge (1959). The latter authors regarded this rock type as a metamorphosed sediment.

If the Sabieboomrante adamellite gneiss is younger than Vioolsdrif Suite rocks then it does not form part of the Richtersveld Subprovince. The contact zone between the two subprovinces therefore passes through the study area, and not farther south as proposed by Colliston and Praekelt (1988).

The Kouefontein granite gneiss has a lateral continuity greater than that of the Sabieboomrante adamellite gneiss. It represents a thick sheet whose position along a significant zone of disconformity is unusual, but not unique. Other granite sheets e.g. Dabbieputs granite gneiss and the leucogranite represent post-tectonic granites intruded probably after tectonism associated with the arc accretion event.

Small plutons of K-granites post-date the Groothoek thrusting phase, transecting lithologies of the Bushmanland and Richtersveld Subprovinces. These plutons differ from typical Richtersveld Suite granites in trace element geochemistry, having high Zr, Nb, Y, but low Sr. They are tentatively correlated with the Spektakel Suite, but may represent a separate suite.

Mafic and ultramafic rocks of the Klipbok complex occupy a significant structural position coincident with the Groothoek Thrust zone, at the contact zone between the Groenrivier suite and Vioolsdrif Granodiorite. Although they occur intermittently along strike, their linear outcrop pattern concordant with lithological contacts suggests that these lithologies have been tectonically disrupted. Geochemically the mafic rocks do not correspond to either arc or mid-oceanic ridge tholeiites and their origin remains unknown. However, taken in context with serpentinites and hornblendites they are interpreted as forming part of an island arc complex which has subsequently been tectonically dismembered along the Groothoek Thrust zone. Ultramafic rocks within the Kouefontein granite gneiss are interpreted as xenolithic fragments caught up during intrusion of the granite.

The controversy over the relationship between the Richtersveld and Bushmanland Subprovinces appears therefore resolved: field and geochemical evidence from the study area reveal that the interface between the two subprovinces is tectonic.

7.2 STRUCTURE

The structural analysis clearly indicates that pre- Pan-African and Pan-African planar and linear structures developed orthogonal to each other. North-northeasterly trending megashears formed during the Late Proterozoic/Early Palaeozoic, segmenting the basement into crustal blocks averaging 4 km in width. The largest and most significant of these shears, the Steenbok Shear, is a major lineament extending from 60 km south of the present area northwards into the Richtersveld.

Most of the shears are rotational faults of displacement ranging between 0,3 and 13,5 km. Some have large strike-slip and dip-slip components in addition to rotational characteristics, indicating an associated complex tectonic history. The Steenbok Shear, for example, has a strike-slip component of 12,4 km and a dip-slip component of 6,3 km. Rotational movement along some shears has steepened strata adjacent to them and formed interference fold structures. This has caused an irregular distribution of supracrustal and intrusive lithologies across the area. An upthrust motion along the shears has offset strata, making correlation difficult. However, a palinspastic reconstruction to Late Proterozoic times, prior to Pan-African deformation, facilitates stratigraphic correlation and explains their irregular distribution. This reconstruction reveals that major shears deflect from their normal strike orientation where they transect large granitic sheets. The latter acted as physical buttresses during Late Proterozoic tectonism, causing a swing in shear zone strike towards the south.

Several north-westerly striking thrust zones, mostly of Early to Mid-Proterozoic age, were identified in the study area. The two most significant are the Ratelfontein and Groothoek Thrusts. The former, characterized by boudinaged quartzite units at the base of the supracrustals of the Richtersveld Subprovince, formed during Early Proterozoic south-westward thrusting. It occurs along a significant zone of lithological and structural discordance; it was reactivated by the Mid-Proterozoic Groothoek Thrust. The latter is a broad zone of ductile deformation at least 2 km wide separating Richtersveld Suite volcano/ plutonic rocks from supracrustals of the Groenrivier suite. It coincides with a discontinuous array of Klipbok complex mafic and ultramafic rocks and magnesian schists which have regional continuity eastwards for at least 150 km. The Groothoek Thrust developed features characteristic of regional thrusting under ductile deformation conditions, for example, a very pronounced regional foliation, re-orientation of linear and fold structures, and rotation of hornblende porphyroblasts in localized zones. The direction of tectonic transport along the thrust was slightly anti-clockwise from that of the Ratelfontein Thrust, as indicated by a well-developed stretching lineation. A minimum displacement of 50 km has been estimated for the Groothoek Thrust zone.

Several other thrust zones were identified in the study area. The most important include the Windvlakte Thrust (oldest thrust in the area), Sabieboomrante, and Chabiesies South Thrust.

A sequence of deformational events and episodes is proposed in an attempt to unravel the sequence of thrusting during the Early to Mid-Proterozoic. The first event (D(R)) is characterized by localized thrusting in supracrustals of the Richtersveld Subprovince (the Windvlakte Thrust). This was followed by development of the Ratelfontein Thrust during the first of several proposed major deformation events, before the Namaqua event (D(N)) at 1200-1100 Ma. The D(N) event is characterized by four episodes during the Mid-Proterozoic. The most significant of these episodes is the second one during which the Groothoek Thrust was formed close to the mafic and ultramafic rocks of the Klipbok complex. Reactivation along this same zone during the third episode formed minor thrusts (e.g., Chabiesies South Thrust). An extensional tectonic phase marks the last of the Namaqua-related episodes.

The Late Proterozoic - Early Palaeozoic (Pan-African) event is comprised of four episodes. During the first episode deformation was initially south-directed. Movement took place along the Stinkfontein Contact, Kromnek and Steenbok Shears, creating a large strike-slip component along these shears. During the second episode south-east-ward directed tectonism caused upthrust movement along all the major shears. The third episode of deformation took place in post-Nama times, involving further left-lateral, transtensional, strike-slip movement along the Steenbok Shear. During this episode Nama sediments were downdropped within the shear in zones of marked change in strike orientation. Rotational movement along the Steenbok and Kromnek Shears deformed the Nama sediments within the shear and those adjacent to the Kouefontein Shear. Brittle deformation structures formed during the last deformation episode. This took place mainly along the Steenbok Shear, at shallower crustal levels.

7.3 METAMORPHISM

The imprints of two metamorphic periods, the Namaqua (1200-1100 Ma) and Pan-African (700-500 Ma) events, are recorded in the study area. These events have obliterated all evidence of assumed pre-Namaqua metamorphism in the area.

The Namaqua metamorphic imprint shows that there is an increase in metamorphic grade in the eastern part of the study area from greenschist facies in the north near the Windvlakte Thrust to upper amphibolite in the extreme south-east. The first metamorphic episode (M(N)1, at 1200 Ma) formed high temperature - low pressure prograde mineral assemblages in metapelites. An inferred ga + st isograd is drawn at the base of the Ratelfontein suite whereas the ga + sil isograd is coincident with the top of the Chabiesies suite, confirming a southward increase in metamorphic grade.

Metamorphism during the Namaqua event continued after a prolonged phase of Groothoek thrusting at 1100 Ma, taking place under conditions of lowering pressure. This establishes that the Namaqua metamorphic event outlasted Groothoek thrusting. Pressure conditions in the eastern part of the area are estimated at 3.5 kbar, based on mineral assemblages in metapelites.

Garnet - biotite geothermometry calculations show that temperatures during the M(N)1 amphibolite facies metamorphic episode ranged between 756° to 778° ± 50°C.

Staurolites in metapelites of the Ratelfontein suite east of Tierkloof Shear contain anomalously high proportions of ZnO (1.24 wt%). Zn became preferential concentrated in staurolite as a result of high temperatures attained during amphibolite facies metamorphism (Namaqua event).

The second episode of metamorphism (M(N)2) is characterized by retrograde assemblages formed in zones of shearing associated with the waning stages of the Namaqua event.

Metamorphism during the Pan-African event comprises four episodes. The first episode (M(P)1, at 700 Ma) was superimposed across a large part of the study area. It reached lower amphibolite facies in the west and greenschist facies in the south-east. The metamorphic grade recorded from the first episode therefore increases towards the west. This interpretation contrasts with that of Joubert (1971) who proposed a regional easterly increase in metamorphic grade in the West Coast Belt. In the south-west of the study area medium pressure-temperature assemblages characterized by kyanite and staurolite formed in basement rocks after the D(P)1 deformation episode. Their development probably resulted from metamorphism brought about by tectonic loading.

Geothermometer calculations on garnet-biotite pairs west of Tierkloof Shear show that temperatures for the first episode ranged between 514° and 630°C ± 50°C. Geobarometer calculations using garnet-plagioclase pairs show that pressures reached 5.5 kbar in the same area.

Retrograde metamorphism during the second episode (M(P)2) took place at about 500 Ma under greenschist facies conditions. This metamorphism is confined to Pan-African shear zones. The third (M(P)3) and fourth (M(P)4) episodes reflect further metamorphism at greenschist facies conditions, associated with phases of reactivation along some of the Pan-African shears, and in particular the Steenbok Shear.

7.4 GEOTECTONIC INTERPRETATION

The Early Proterozoic geotectonic evolution in the north-western part of the Namaqua Province commenced with the formation of an arc system. Thereafter an arc accretion event took place at about 1850 Ma. The Richtersveld Subprovince comprising part of an island arc complex was juxtaposed against a terrane with continentally-derived strata of the Bushmanland Subprovince. The suture is marked by the Ratelfontein Thrust, which forms a broad zone of imbricate thrusting above boudinaged quartzites of the Ratelfontein suite. Mafic and ultramafic rocks of the Klipbok complex were emplaced during the arc accretion event.

A south-westward propagated thrusting and magmatic accretion model is proposed to account for post arc accretion and pre-Namaqua deformation in the westernmost part of the Namaqua Province. Major thrust zones developed into the foreland, dislocation taking place along metamorphic interfaces in the lithosphere. Four events are proposed to explain deformation and emplacement of granitic sheets in the broader region of the north-western Namaqua Province, although evidence for all the events cannot be confirmed in the study area.

Mid-Proterozoic tectonism is characterized by southerly movement along the Groothoek Thrust, during the latter part of the Namaqua metamorphic event. Dislocation was focussed along the base of competent Vioolsdrif Suite volcanic/plutonic rocks coinciding with mafic and ultramafic rocks of the Klipbok complex. Movement of the Richtersveld Subprovince rocks southward over those of the Bushmanland Subprovince took place slightly prior to 1100 Ma, the Namaqua metamorphic event outlasting Groothoek thrusting.

Reactivation during the Mid-Proterozoic was focussed along minor thrusts (e.g., Chabiesies South thrust) just south of the Ratelfontein Thrust, during the waning stages of the Namaqua event.

The last Namaqua- related episode formed in response to northward extension, developing brittle deformation structures.

A model of combined shearing and thrusting adequately explains initial left-lateral movement along the Stinkfontein Contact, Kromnek and Steenbok Shears, during the early part of the Pan-African event. Tectonism was accompanied by south-easterly movement of nappe structures over basement rocks in the foreland region of the Gariep Belt. This tectonic activity effected the first Pan-African metamorphic episode (M(P)1), through tectonic loading. Subsequent compression changed the orientation of the stress field towards the south-east, causing westerly upthrusting of crustal blocks along steeply-dipping shears, under greenschist facies conditions. This formed a recess area near Eksteenfontein, where crustal blocks became rotated, especially where the shears transected large granite sheets.

Post Nama tectonism was south-easterly directed and was concentrated mainly along the Steenbok and Kouefontein Shears. Nama strata were uplifted and eroded west of Steenbok Shear and downdropped adjacent to the Kouefontein Shear. Left-lateral, transtensional movement along the Steenbok Shear caused Nama sediments to be downdropped and deformed within the shear. The last movement along the Steenbok Shear took place at shallower crustal levels, forming brittle deformation structures during lower greenschist facies metamorphism.

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APPENDIX A-1 METHODS OF STUDY

A-1.1 Field Mapping

The writer spent an initial period of six months in the field, from July to December 1983, whilst on sabbatical leave from the University of Port Elizabeth. A further period of one month during the 1985 July vacation was needed for additional detailed mapping and sample collecting.

Mapping was carried out directly on to aerial photographs at a scale of 1:36,000. A fairly dense network of observation stations could be achieved owing to the generally good outcrop (about 60%) over most of the area. Each station was readily identified on photo pairs, with the aid of a pocket stereoscope, and the relevant structural and sampling data recorded opposite the corresponding number in a field note book. Care was taken with regard to collecting oriented specimens in the field, especially where subsequent detailed structural studies were to be carried out.

A-1.2 Map Compilation

All station numbers recorded on each aerial photograph were transferred on to the appropriate 1:50,000 topocadastral maps with the aid of a Sketchmaster. The main geological contacts and limits of outcrop were also drawn in at this stage. A photographic mosaic at a scale of 1:25,000 was compiled of all these data points so that more detail could be inserted at this enlarged scale. This scale was also more appropriate for locating rock sample numbers and carrying out a structural analysis over the entire area.

Once all the relevant geological detail had been plotted the map was photoreduced to the final scale of 1:50,000 (see Annexure 1).

A preliminary photo-interpretation was carried out before field work commenced, but was found to be of limited use during the mapping programme, especially regarding the location of lithological contacts. A subsequent photo-interpretation, largely delineating foliation trends and younger structures, was carried out by transferring this structural data from the aerial photographs via a Sketchmaster on to 1:50,000 topocadastral sheets. This map was photo-reduced to a scale of 1:165,000, and appears as Fig. 4.1.

A portion of the north-east corner of the area under investigation was mapped on a scale of 1:15,000, to provide more detail on the distribution of mafic, ultramafic and magnesian rocks, in the Groothoek Thrust zone. This map was photo-reduced to a scale of 1:30,000 (see Annexure 2).

A-1.3 Analysis of structural data.

Structural data (poles to foliations) were plotted on an equal-area Schmidt net, and contoured to reveal density of plots according to the methods suggested by Ragan (1968). This data, together with lineation plots, is presented in Annexure 3. A separate diagram (Fig. 4.10) shows lineation trends over the area investigated their lengths plotted as the cosine of the angle of lineation plunge.

A-1.4 X-ray Analyses.

X-ray diffraction studies on unknown minerals, were carried out on a Siemens 500 diffractometer fitted with a DACO microprocessor, at the University of Port Elizabeth.

The operating conditions were as follows:-

Co tube run at 40 kV and 30 Ma.
Aperture slits 1, 2, and 3 = 1°, detector slit = 0.05°.
Step-scan interval 0.05°.

Samples were prepared by crushing to - 200 mesh in a McCrone mill. Each sample was mounted in a Siemens plastic holder, and levelled with a glass slide. This might give rise to preferred orientations, but the 2 u grain size is thought to largely obviate this problem.

d-Spacings were computed directly by the DACO microprocessor. The system was previously tested against a standard quartz sample from which the following d-spacings were obtained:-

4.2500; 3.3378; 2.4547; 2.2729; 2.1260; 1.9784; 1.8161.

The d-spacings were matched against the ASTM index (Published by the Joint Committee on Powder Diffraction Standards, 1974) and the unknowns were identified.

A-1.5 Electron microprobe mineral analyses

A JEOL 733 electron microprobe in the Geology Department, Rhodes University, Grahamstown, was used to carry out mineral analyses. Thin sections were cut to approximately twice standard thickness and final polish accomplished with a 1/4 micron paste on a Durr cloth. Each sample was coated with carbon to aid conduction.

The operating conditions were as follows:-
Beam current: 25 nA
Accelerating voltage: 15 Kv
Analysing crystals: TAP (Si, Al, Mg, Na)
(Flow counters with Argon/Methane 90:10)
Quartz (LiF for Fe, Mn; PET for Cr, Ca, Ti, K)
(Sealed counters)

Garnet, plagioclase and staurolite were analysed with a 10 u diameter beam, but for biotite the beam was optimised to 1-2 u. Natural and synthetic standards were used. The data was processed by the ZAF correction program. All Fe is reported as FeO.

Mineral analyses are shown in Appendices C-1 to C-4.

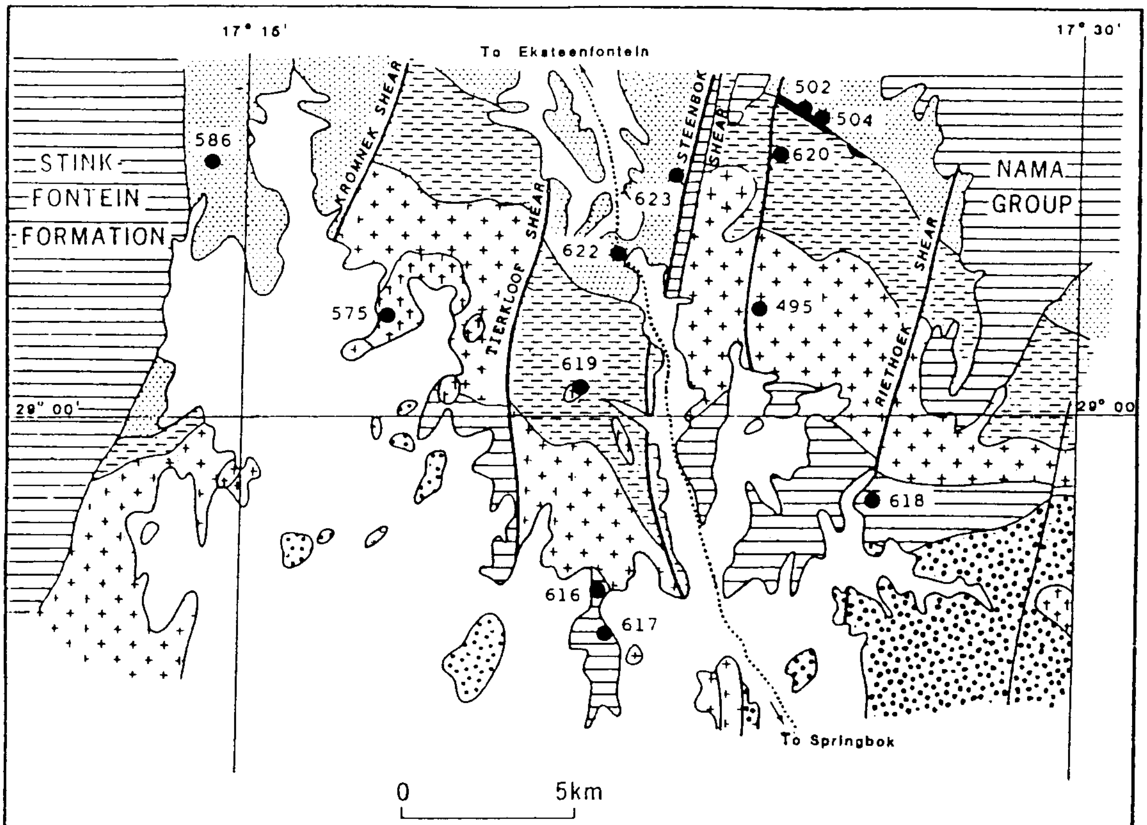
A-1.6 Geochemical analyses

Geochemical analyses were carried out by Dr D.L. Reid and Dr J.M. Moore in the Department of Geochemistry, at the University of Cape Town. All major and trace elements were analysed by XRF using standard methods described by Duncan et al. (1984a; In: Erlank, A. J. (Ed.) Petrogenesis of the volcanic rocks of the Karoo Province Spec. Publ. geol. Soc. S. Afr., **13**, 389-395).

The results are presented in Tables 3.6, 3.21, 3.24, 3.26 and 3.27.

A locality map of geochemical sample sites is included as Appendix B-1.

Appendix B - 1 Locality map, geochemical sample sites



SIMPLIFIED GEOLOGICAL MAP OF AN AREA 30km SOUTH-EAST OF EKSTEENFONTEIN

LEGEND

Richtersveld Subprovince:
- Windvlakte suite; Vioolsdrif Suite
- Groenrivier, Groothoek, Ratelfontein suites (= "Transition zone")
- Klipbok complex (mafic/ultramafic rocks)

Bushmanland Subprovince:
- Kouefontein granite gneiss
- Sabieboomrante adamellite gneiss
- Chabiesies suite; Gladkop Suite

- Kromnek-type granites

SAMPLE No.	ROCK TYPE
PB 495	ultramafic
PB 502	amphibolite (Klipbok complex)
PB 504	ultramafic "
PB 575	granite (Kromnek)
PB 586	megacrystic granite
PB 616	adamellite gneiss (Skurwehoogte)
PB 617	biotite granite (Skurwehoogte)
PB 618	adamellite gneiss (Sabieboomrante)
PB 619	biotite granite (Donkiehel)
PB 620	leucogranite
PB 622	granodiorite (Kliphoogte)
PB 623	granodiorite (3 km NE of Kliphoogte)

APPENDIX C-1 Microprobe Analyses of Staurolite

Sample No	592B St 1	592B St 1q	592B St 2	593A St 1	593A St 2
Domain No	1	1	1	1	1
SiO2	28.93	27.46	28.06	28.76	28.32
TiO2	0.47	0.52	0.46	0.42	0.44
Al2O3	53.56	55.12	52.75	53.27	53.32
FeO	13.03	14.10	14.25	14.23	14.05
ZnO	0.38	0.40	0.25	0.37	0.20
MnO	0.01	0.00	0.02	0.00	0.00
MgO	1.77	1.49	1.45	1.13	1.19
Total	98.15	99.09	97.24	98.18	97.52
	Cations per 46 Oxygens				
Si	7.963	7.542	7.852	7.959	7.882
Ti	0.097	0.107	0.096	0.087	0.092
Al	17.381	17.847	17.403	17.380	17.496
Fe	2.999	3.238	3.335	3.293	3.270
Zn	0.077	0.081	0.051	0.075	0.041
Mn	0.002	0.000	0.004	0.000	0.000
Mg	0.726	0.609	0.604	0.466	0.493
Total	29.248	29.426	29.348	29.262	29.276

APPENDIX C-1 Microprobe Analyses of Staurolite (cont.)

Sample No	475A St 1	475A St 2	475A St 3	475A St 4	476 St 1	476 St 2
Domain No	2	2	2	2	2	2
SiO2	28.60	28.47	28.63	28.20	28.24	28.24
TiO2	0.45	0.43	0.42	0.32	0.50	0.53
Al2O3	52.78	53.48	53.48	53.50	53.11	52.77
FeO	14.97	14.51	14.90	15.09	14.74	15.18
ZnO	0.00	0.12	0.11	0.13	0.12	0.14
MnO	0.01	0.01	0.01	0.01	0.01	0.01
MgO	1.48	1.38	1.55	1.38	1.32	1.50
Total	98.29	98.40	99.10	98.63	98.04	98.37
	Cations per 46 Oxygens					
Si	7.927	7.868	7.870	7.801	7.847	7.842
Ti	0.093	0.089	0.086	0.066	0.104	0.110
Al	17.248	17.424	17.332	17.449	17.398	17.277
Fe	3.470	3.353	3.425	3.491	3.425	3.525
Zn	0.000	0.024	0.022	0.026	0.024	0.028
Mn	0.002	0.002	0.002	0.002	0.002	0.002
Mg	0.611	0.568	0.635	0.569	0.546	0.620
Total	29.354	29.330	29.375	29.407	29.349	29.408

APPENDIX C-1 Microprobe Analyses of Staurolite (cont.)

Sample No	317A St 1	317A St 2	317B St 1	317B St 2	317B St 3	317E St 1	317E St 2
Domain No	4	4	4	4	4	4	4
SiO2	28.01	28.16	28.11	28.66	28.50	28.10	28.29
TiO2	0.66	0.55	0.67	0.48	0.49	0.57	0.54
Al2O3	54.39	54.78	54.80	55.26	55.49	54.98	55.28
FeO	12.35	12.45	11.88	11.70	11.73	12.19	11.63
ZnO	1.00	1.19	1.22	1.38	1.44	1.17	1.28
MnO	0.07	0.09	0.08	0.10	0.09	0.09	0.07
MgO	1.17	1.17	1.03	1.05	1.01	1.08	0.97
Total	97.65	98.39	97.79	98.63	98.75	98.18	98.06
				Cations per 46 Oxygens			
Si	7.758	7.750	7.762	7.837	7.788	7.738	7.777
Ti	0.137	0.113	0.139	0.098	0.100	0.118	0.111
Al	17.761	17.775	17.840	17.815	17.878	17.850	17.916
Fe	2.861	2.865	2.743	2.675	2.681	2.807	2.673
Zn	0.204	0.241	0.248	0.278	0.290	0.237	0.259
Mn	0.016	0.020	0.018	0.023	0.020	0.020	0.016
Mg	0.483	0.479	0.423	0.427	0.411	0.443	0.397
Total	29.222	29.248	29.177	29.156	29.171	29.217	29.152

APPENDIX C-2 Microprobe Analyses of Garnet

Sample No	592D Ga 1 Core	592D Ga 1 Rim	592D Ga 2 Core	592D Ga 2 Rim	592D Ga 3 Core	592D Ga 3 Rim	593A Ga 1 Core	593A Ga 1 Rim
Domain No	1	1	1	1	1	1	1	1
SiO2	37.32	36.91	37.72	37.41	37.22	37.91	36.91	37.14
TiO2	0.00	0.00	0.05	0.01	0.22	0.03	0.01	0.01
Al2O3	21.35	21.25	21.15	21.35	21.29	21.32	21.43	21.37
FeO	34.83	35.67	34.30	34.66	34.35	34.69	37.51	37.17
MnO	0.19	0.17	0.20	0.39	1.13	0.30	0.23	0.10
MgO	2.16	2.22	2.10	2.18	2.09	2.29	1.76	2.02
CaO	3.75	3.32	3.83	3.63	3.35	3.25	1.89	1.91
Na2O	0.00	0.00	0.02	0.00	0.00	0.00	0.01	0.00
Total	99.60	99.54	99.37	99.63	99.65	99.79	99.75	99.72
				Cations per 24 Oxygens				
Si	6.014	5.975	6.074	6.023	6.002	6.076	5.987	6.010
Ti	0.000	0.000	0.006	0.001	0.026	0.003	0.001	0.001
Al	4.056	4.055	4.015	4.053	4.048	4.028	4.098	4.076
Fe	4.694	4.829	4.619	4.667	4.633	4.650	5.088	5.030
Mn	0.025	0.023	0.027	0.053	0.154	0.040	0.031	0.013
Mg	0.518	0.535	0.504	0.523	0.502	0.547	0.425	0.487
Ca	0.647	0.575	0.660	0.626	0.578	0.558	0.328	0.331
Na	0.000	0.000	0.006	0.000	0.000	0.000	0.003	0.000
Total	15.957	15.996	15.914	15.948	15.946	15.905	15.964	15.950
Alm	79.746	80.973	79.488	79.514	78.945	80.229	86.626	85.807
Spt	0.440	0.390	0.468	0.904	2.630	0.702	0.537	0.233
Pyr	8.812	8.980	8.672	8.912	8.559	9.437	7.243	8.309
Gro	11.000	9.655	11.371	10.669	9.864	9.630	5.592	5.649

APPENDIX C-2 Microprobe Analyses of Garnet (cont.)

Sample No	593A Ga 2 Core	593A Ga 2 Rim	593A Ga 3 Core	593A Ga 3 Rim	593A Ga 4 Core	593A Ga 4 Rim	475A Ga 1 Core	475A Ga 1 Rim
Domain No	1	1	1	1	1	1	2	2
SiO2	37.46	37.33	36.83	36.85	37.14	37.35	37.17	37.41
TiO2	0.04	0.01	0.02	0.03	0.00	0.01	0.00	0.00
Al2O3	21.13	21.36	21.67	21.75	21.41	21.57	21.33	21.83
FeO	35.45	37.72	35.86	36.74	37.35	37.44	35.69	35.99
MnO	0.60	0.13	0.39	0.21	0.19	0.03	1.47	1.25
MgO	1.17	1.80	1.39	1.90	1.66	1.99	2.58	2.90
CaO	3.99	1.95	3.32	2.62	1.84	2.13	1.21	1.24
Na2O	0.01	0.02	0.00	0.00	0.00	0.01	0.00	0.01
Total	99.85	100.32	99.48	100.10	99.59	100.53	99.45	100.73
			Cations per 24 Oxygens					
Si	6.052	6.017	5.974	5.945	6.023	5.997	6.015	5.972
Ti	0.004	0.001	0.002	0.003	0.000	0.001	0.000	0.000
Al	4.025	4.059	4.143	4.137	4.093	4.083	4.069	4.108
Fe	4.790	5.084	4.864	4.957	5.066	5.028	4.830	4.805
Mn	0.082	0.017	0.053	0.028	0.026	0.004	0.201	0.182
Mg	0.281	0.432	0.336	0.456	0.401	0.476	0.622	0.690
Ca	0.690	0.336	0.577	0.452	0.319	0.366	0.209	0.212
Na	0.003	0.006	0.000	0.000	0.000	0.003	0.000	0.003
Total	15.931	15.955	15.951	15.982	15.929	15.960	15.949	15.974
Alm	81.956	86.598	83.424	84.083	87.148	85.587	82.374	81.584
Spt	1.404	0.301	0.917	0.485	0.449	0.068	3.436	3.099
Pyr	4.820	7.364	5.762	7.748	6.902	8.106	10.611	11.714
Gro	11.818	5.735	9.895	7.682	5.500	6.238	3.578	3.601

APPENDIX C-2 Microprobe Analyses of Garnet (cont.)

Sample No	475A Ga 2 Core	475A Ga 2 Rim	475A Ga 3 Core	475A Ga 3 Rim	475A Ga 4 Core	475A Ga 4 Rim
Domain No	2	2	2	2	2	2
SiO2	37.85	36.94	37.73	37.65	37.78	37.63
TiO2	0.00	0.01	0.00	0.00	0.00	0.00
Al2O3	21.01	21.51	21.17	21.10	21.24	21.25
FeO	35.88	36.12	35.14	35.68	35.40	35.63
MnO	1.48	1.35	1.24	1.36	1.29	1.50
MgO	2.23	2.53	3.16	2.87	3.19	2.30
CaO	1.19	1.25	1.23	1.19	1.21	1.25
Na2O	0.00	0.00	0.00	0.00	0.00	0.00
Total	99.64	99.71	99.67	99.85	100.11	99.56
	Cations per 24 Oxygens					
Si	6.108	5.974	6.062	6.058	6.050	6.074
Ti	0.000	0.001	0.000	0.000	0.000	0.000
Al	3.997	4.101	4.010	4.002	4.009	4.044
Fe	4.842	4.885	4.722	4.801	4.741	4.810
Mn	0.202	0.184	0.168	0.185	0.175	0.205
Mg	0.536	0.609	0.756	0.688	0.761	0.553
Ca	0.205	0.216	0.211	0.205	0.207	0.216
Na	0.000	0.000	0.000	0.000	0.000	0.000
Total	15.892	15.973	15.932	15.940	15.945	15.903
Alm	83.680	82.849	80.591	81.654	80.561	83.151
Spt	3.496	3.136	2.880	3.152	2.973	3.545
Pyr	9.267	10.341	12.914	11.704	12.936	9.565
Gro	3.555	3.673	3.614	3.489	3.528	3.737

APPENDIX C-2 Microprobe Analyses of Garnet (cont.)

Sample No	475A Ga 5 Core	475A Ga 5 Rim	475A Ga 6 Core	475A Ga 6 Rim	476 Ga 1 Core	476 Ga 1 Rim	476 Ga 2 Core	476 Ga 2 Rim
Domain No	2	2	2	2	2	2	2	2
SiO2	37.67	37.42	37.53	37.37	37.71	37.49	37.29	37.64
TiO2	0.00	0.01	0.00	0.00	0.00	0.00	0.00	0.00
Al2O3	21.46	21.28	21.17	21.43	21.71	21.66	21.43	21.70
FeO	34.84	34.98	35.27	35.27	35.32	36.08	35.37	35.41
MnO	1.21	1.15	1.13	1.07	1.28	1.17	1.33	1.29
MgO	3.38	3.45	3.24	3.44	3.43	2.91	2.96	2.85
CaO	1.24	1.19	1.26	1.23	1.19	1.30	1.20	1.23
Na2O	0.00	0.00	0.00	0.01	0.00	0.00	0.00	0.00
Total	99.80	99.48	99.60	99.82	100.64	100.61	99.58	100.12
Cations per 24 Oxygens								
Si	6.034	6.022	6.039	5.999	6.000	5.991	6.011	6.025
Ti	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
Al	4.052	4.037	4.016	4.055	4.072	4.080	4.072	4.095
Fe	4.667	4.708	4.747	4.735	4.700	4.822	4.768	4.740
Mn	0.164	0.156	0.154	0.145	0.172	0.158	0.181	0.174
Mg	0.806	0.827	0.777	0.823	0.813	0.693	0.711	0.679
Ca	0.212	0.205	0.217	0.211	0.202	0.222	0.207	0.211
Na	0.000	0.000	0.000	0.004	0.000	0.000	0.000	0.000
Total	15.939	15.958	15.951	15.974	15.962	15.968	15.952	15.927
Alm	79.766	79.831	80.520	80.050	79.813	81.784	81.255	81.644
Spt	2.805	2.658	2.612	2.459	2.929	2.686	3.094	3.012
Pyr	13.790	14.030	13.181	13.913	13.811	11.754	12.117	11.709
Gro	3.637	3.479	3.685	3.576	3.445	3.775	3.532	3.633

APPENDIX C-2 Microprobe Analyses of Garnet (cont.)

Sample No	476 Ga 3 Core	476 Ga 3 Rim	476 Ga 4 Core	476 Ga 4 Rim	476 Ga 5 Core	476 Ga 5 Rim	476 Ga 6 Core	476 Ga 6 Rim
Domain No	2	2	2	2	2	2	2	2
SiO2	37.60	37.63	37.62	37.54	37.07	37.40	37.22	37.19
TiO2	0.01	0.01	0.00	0.01	0.00	0.00	0.00	0.00
Al2O3	21.63	21.95	21.77	21.69	21.31	21.44	21.59	21.39
FeO	34.59	35.23	34.73	35.28	34.93	35.58	35.75	35.40
MnO	1.13	1.45	1.12	1.18	1.20	1.31	1.20	1.23
MgO	3.50	3.08	3.40	3.10	3.41	3.36	3.20	2.94
CaO	1.19	1.24	1.14	1.24	1.20	1.26	1.25	1.28
Na2O	0.01	0.02	0.01	0.02	0.00	0.01	0.01	0.01
Total	99.66	100.61	99.79	100.06	99.12	100.36	100.22	99.44
				Cations per 24 Oxygens				
Si	6.021	5.992	6.018	6.009	5.994	5.986	5.969	6.005
Ti	0.001	0.001	0.000	0.001	0.000	0.000	0.000	0.000
Al	4.084	4.120	4.105	4.093	4.062	4.045	4.081	4.072
Fe	4.633	4.691	4.646	4.723	4.723	4.762	4.794	4.781
Mn	0.153	0.195	0.151	0.160	0.164	0.177	0.163	0.168
Mg	0.835	0.730	0.810	0.739	0.821	0.801	0.764	0.707
Ca	0.204	0.211	0.195	0.212	0.207	0.216	0.214	0.221
Na	0.003	0.006	0.003	0.006	0.000	0.003	0.003	0.003
Total	15.936	15.949	15.930	15.945	15.974	15.992	15.991	15.959
Alm	79.524	80.478	80.053	80.939	79.822	79.939	80.755	81.333
Spt	2.631	3.354	2.614	2.741	2.777	2.981	2.745	2.862
Pyr	14.339	12.537	13.965	12.673	13.886	13.452	12.881	12.037
Gro	3.505	3.629	3.366	3.644	3.513	3.627	3.617	3.767

APPENDIX C-2 Microprobe Analyses of Garnet (cont.)

Sample No	317A Ga 1 Core	317A Ga 1 Rim	317A Ga 2 Core	317A Ga 2 Rim	317A Ga 3 Core	317A Ga 3 Rim	317A Ga 4 Core	317A Ga 4 Rim
Domain No	4	4	4	4	4	4	4	4
SiO2	37.82	38.08	37.44	37.60	37.44	37.49	37.54	37.47
TiO2	0.03	0.00	0.03	0.01	0.01	0.00	0.00	0.00
Al2O3	21.30	21.58	21.33	21.36	21.59	21.76	21.54	21.93
FeO	35.93	34.84	36.07	35.27	35.54	34.84	35.18	34.52
MnO	0.51	0.25	0.78	0.71	1.38	0.27	0.92	0.31
MgO	4.25	5.06	4.16	4.53	3.94	4.94	4.28	4.91
CaO	0.42	0.41	0.35	0.37	0.53	0.39	0.60	0.42
Na2O	0.00	0.00	0.00	0.02	0.00	0.01	0.01	0.00
Total	100.26	100.22	100.16	99.87	100.43	99.70	100.07	99.56
Cations per 24 Oxygens								
Si	6.026	6.026	5.988	6.005	5.974	5.974	5.990	5.970
Ti	0.003	0.000	0.003	0.001	0.001	0.000	0.000	0.000
Al	4.001	4.026	4.022	4.022	4.061	4.087	4.052	4.119
Fe	4.788	4.611	4.825	4.711	4.742	4.643	4.695	4.600
Mn	0.068	0.033	0.105	0.096	0.186	0.036	0.124	0.041
Mg	1.009	1.193	0.991	1.078	0.937	1.173	1.017	1.166
Ca	0.071	0.069	0.059	0.063	0.090	0.066	0.102	0.071
Na	0.000	0.000	0.000	0.006	0.000	0.001	0.001	0.000
Total	15.969	15.960	15.996	15.984	15.994	15.982	15.984	15.969
Alm	80.636	78.055	80.655	79.196	79.618	78.441	79.042	78.238
Spt	1.158	0.567	1.766	1.613	3.131	0.615	2.093	0.710
Pyr	16.997	20.201	16.576	18.126	15.729	19.819	17.136	19.830
Gro	1.207	1.176	1.001	1.064	1.520	1.123	1.727	1.219

APPENDIX C-2 Microprobe Analyses of Garnet (cont.)

Sample No	317A Ga 5 Core	317A Ga 5 Rim	317A Ga 6 Core	317A Ga 6 Rim	462Y Ga 1 Core	462Y Ga 1 Rim	462Y Ga 2 Core	462Y Ga 2 Rim
Domain No	4	4	4	4	5	5	5	5
SiO2	37.46	37.70	38.15	37.92	37.56	37.65	37.65	37.11
TiO2	0.00	0.00	0.00	0.02	0.02	0.01	0.00	0.00
Al2O3	21.54	21.78	21.51	21.49	21.63	21.59	21.30	21.47
FeO	35.35	34.99	34.68	34.58	33.61	33.03	31.95	31.33
MnO	1.03	0.34	0.30	0.25	4.13	5.06	5.79	6.81
MgO	4.13	4.88	4.93	4.95	2.56	2.34	2.07	1.72
CaO	0.48	0.45	0.33	0.34	1.02	1.03	1.08	1.10
Na2O	0.00	0.01	0.00	0.00	0.00	0.00	0.00	0.00
Total	99.99	100.15	99.90	99.55	100.53	100.71	99.84	99.54
				Cations per 24 Oxygens				
Si	5.988	5.982	6.051	6.037	6.013	6.024	6.071	6.021
Ti	0.000	0.000	0.000	0.002	0.002	0.001	0.000	0.000
Al	4.059	4.074	4.022	4.033	4.082	4.072	4.049	4.107
Fe	4.726	4.644	4.600	4.604	4.499	4.419	4.308	4.251
Mn	0.139	0.045	0.040	0.033	0.560	0.685	0.790	0.936
Mg	0.984	1.154	1.165	1.174	0.610	0.558	0.497	0.416
Ca	0.082	0.076	0.056	0.058	0.175	0.176	0.186	0.191
Na	0.001	0.003	0.000	0.000	0.000	0.000	0.000	0.000
Total	15.981	15.981	15.936	15.943	15.943	15.938	15.904	15.924
Alm	79.675	78.440	78.477	78.431	76.978	75.679	74.498	73.369
Spt	2.351	0.771	0.687	0.574	9.580	11.742	13.673	16.152
Pyr	16.587	19.495	19.880	20.006	10.448	9.554	8.601	7.177
Gro	1.385	1.292	0.955	0.988	2.993	3.023	3.226	3.300

APPENDIX C-2 Microprobe Analyses of Garnet (cont.)

Sample No	462Y Ga 3 Core	462Y Ga 3 Rim	462Y Ga 4 Core	462Y Ga 4 Rim	462Y Ga 5 Core	462Y Ga 5 Rim	462Y Ga 6 Core	462Y Ga 6 Rim
Domain No	5	5	5	5	5	5	5	5
SiO2	37.59	37.21	37.31	37.57	37.22	37.01	37.28	37.70
TiO2	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.01
Al2O3	21.18	21.31	21.27	21.26	21.23	21.47	21.57	21.28
FeO	32.33	31.99	33.78	33.60	33.04	32.45	34.23	33.48
MnO	6.29	6.52	3.35	4.32	5.87	5.74	3.28	4.02
MgO	2.00	1.82	2.77	2.39	2.02	1.99	2.86	2.48
CaO	0.96	1.00	1.00	1.03	0.93	0.96	1.01	1.07
Na2O	0.00	0.03	0.01	0.02	0.00	0.00	0.02	0.01
Total	100.35	99.88	99.49	100.19	100.31	99.62	100.25	100.05
Cations per 24 Oxygens								
Si	6.054	6.026	6.028	6.044	6.011	6.002	5.985	6.060
Ti	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.001
Al	4.021	4.068	4.051	4.032	4.042	4.105	4.083	4.032
Fe	4.354	4.332	4.564	4.520	4.463	4.401	4.596	4.500
Mn	0.858	0.894	0.458	0.588	0.803	0.788	0.446	0.547
Mg	0.480	0.439	0.667	0.573	0.486	0.481	0.684	0.594
Ca	0.165	0.173	0.173	0.177	0.160	0.166	0.173	0.184
Na	0.000	0.009	0.003	0.006	0.000	0.000	0.006	0.003
Total	15.934	15.944	15.947	15.942	15.967	15.945	15.975	15.923
Alm	74.331	74.191	77.851	77.145	75.474	75.395	77.897	77.246
Spt	14.647	15.315	7.819	10.046	13.581	13.507	7.560	9.394
Pyr	8.194	7.521	11.376	9.778	8.222	8.239	11.598	10.196
Gro	2.827	2.971	2.952	3.029	2.721	2.857	2.944	3.163

APPENDIX C-2 Microprobe Analyses of Garnet (cont.)

Sample No	8978 Ga 1 Core	8978 Ga 1 Rim	8978 Ga 2 Core	8978 Ga 2 Rim	8978 Ga 3 Core	8978 Ga 3 Rim
Domain No	5	5	5	5	5	5
SiO_2	37.57	37.67	37.64	37.36	37.47	37.41
TiO_2	0.00	0.00	0.00	0.00	0.00	0.00
Al_2O_3	21.41	21.30	21.47	21.38	21.34	21.23
FeO	31.03	28.91	31.35	30.95	29.29	29.23
MnO	5.88	8.77	5.65	6.30	8.51	8.66
MgO	2.56	2.15	2.49	2.38	1.99	1.87
CaO	1.08	1.01	1.13	0.97	0.90	1.03
Na_2O	0.01	0.01	0.01	0.01	0.00	0.00
Total	99.54	99.82	99.74	99.35	99.50	99.43
Cations per 24 Oxygens						
Si	6.055	6.071	6.055	6.044	6.063	6.065
Ti	0.000	0.000	0.000	0.000	0.000	0.000
Al	4.068	4.047	4.072	4.077	4.071	4.057
Fe	4.182	3.896	4.218	4.187	3.964	3.963
Mn	0.802	1.197	0.769	0.863	1.166	1.189
Mg	0.614	0.516	0.597	0.573	0.479	0.451
Ca	0.186	0.174	0.194	0.168	0.156	0.178
Na	0.003	0.003	0.003	0.003	0.000	0.000
Total	15.912	15.906	15.910	15.918	15.901	15.906
Alm	72.278	67.361	72.979	72.288	68.742	68.529
Spt	13.872	20.696	13.321	14.903	20.228	20.563
Pyr	10.626	8.927	10.329	9.905	8.322	7.812
Gro	3.223	3.015	3.370	2.902	2.706	3.093

APPENDIX C-3 Microprobe Analyses of Biotite

Sample No	592D Bt 1	592D Bt 2	592D Bt 3	593A Bt 1	593A Bt 2	475A Bt 1	475A Bt 2	475A Bt 3
Domain No	1	1	1	1	1	2	2	2
SiO2	36.97	36.18	36.91	36.21	36.50	36.96	36.04	36.61
TiO2	1.50	1.41	1.41	1.60	1.51	1.18	1.54	1.61
Al2O3	19.83	19.64	19.86	20.11	20.01	19.72	19.16	18.62
FeO	21.63	20.38	20.99	23.04	23.14	21.65	21.37	22.04
MnO	0.00	0.00	0.03	0.00	0.00	0.00	0.00	0.00
MgO	9.87	9.53	9.86	7.84	8.22	8.61	9.39	9.79
CaO	0.00	0.00	0.11	0.00	0.00	0.02	0.00	0.00
Na2O	0.21	0.19	0.16	0.24	0.30	0.15	0.18	0.12
K2O	7.80	8.47	7.55	7.50	7.34	8.50	8.82	8.35
Total	97.81	95.80	96.88	96.54	97.02	96.79	96.50	97.14

Cations per 24 Oxygens

Si	5.945	5.945	5.968	5.937	5.949	6.030	5.926	5.975
Ti	0.181	0.174	0.171	0.197	0.185	0.144	0.190	0.197
Al	3.759	3.804	3.785	3.887	3.845	3.793	3.714	3.583
Fe	2.909	2.800	2.838	3.159	3.154	2.954	2.939	3.008
Mn	0.000	0.000	0.004	0.000	0.000	0.000	0.000	0.000
Mg	2.365	2.333	2.376	1.916	1.996	2.093	2.301	2.381
Ca	0.000	0.000	0.019	0.000	0.000	0.003	0.000	0.000
Na	0.065	0.060	0.050	0.076	0.094	0.047	0.057	0.037
K	1.600	1.775	1.557	1.569	1.526	1.769	1.850	1.738
Total	16.826	16.895	16.771	16.743	16.753	16.836	16.979	16.923

APPENDIX C-3 Microprobe Analyses of Biotite (cont.)

Sample No	475A Bt 4	475A Bt 5	476 Bt 1	476 Bt 2	476 Bt 3	317A Bt 1	317A Bt 2	317A Bt 3
Domain No	2	2	2	2	2	4	4	4
SiO2	35.88	35.41	36.02	36.09	36.59	35.43	36.74	37.13
TiO2	1.58	1.56	1.44	1.48	1.61	1.01	1.88	1.15
Al2O3	19.02	19.24	18.91	18.98	19.11	18.90	18.86	19.09
FeO	21.89	23.26	22.55	22.61	22.59	22.21	22.08	20.95
MnO	0.02	0.01	0.02	0.00	0.03	0.12	0.08	0.12
MgO	9.48	9.31	9.23	9.37	9.10	11.40	10.17	10.53
CaO	0.01	0.15	0.00	0.00	0.00	0.00	0.01	0.00
Na2O	0.23	0.20	0.12	0.18	0.12	0.07	0.15	0.11
K2O	8.68	7.96	8.12	7.90	7.12	6.88	7.53	6.93
Total	96.79	97.10	96.41	96.61	96.27	96.02	97.50	96.01
	Cations per 24 Oxygens							
Si	5.898	5.822	5.938	5.930	5.992	5.827	5.945	6.037
Ti	0.195	0.192	0.178	0.182	0.198	0.124	0.228	0.140
Al	3.686	3.729	3.675	3.676	3.690	3.664	3.598	3.659
Fe	3.009	3.198	3.109	3.107	3.094	3.055	2.988	2.849
Mn	0.002	0.001	0.002	0.000	0.004	0.016	0.010	0.016
Mg	2.322	2.281	2.267	2.294	2.221	2.794	2.452	2.551
Ca	0.001	0.026	0.000	0.000	0.000	0.000	0.001	0.000
Na	0.073	0.063	0.038	0.057	0.038	0.022	0.047	0.034
K	1.820	1.669	1.707	1.656	1.487	1.443	1.554	1.437
Total	17.010	16.986	16.918	16.905	16.726	16.948	16.827	16.727

APPENDIX C-3 Microprobe Analyses of Biotite (cont.)

Sample No	317A Bt 4	317A Bt 5	317A Bt 6	462Y Bt 1	462Y Bt 2	462Y Bt 3	462Y Bt 4	462Y Bt 5
Domain No	4	4	4	5	5	5	5	5
SiO2	35.72	35.03	35.83	34.31	34.23	33.87	34.19	35.95
TiO2	1.25	1.24	1.39	2.20	2.24	2.19	2.16	2.36
Al2O3	19.19	19.42	19.10	18.57	18.72	18.65	18.41	19.15
FeO	21.86	21.43	20.90	26.27	27.09	27.22	27.25	25.23
MnO	0.14	0.13	0.12	0.25	0.24	0.22	0.26	0.22
MgO	11.09	11.10	10.97	5.57	5.82	5.69	5.93	5.70
CaO	0.00	0.00	0.04	0.00	0.00	0.00	0.00	0.00
Na2O	0.09	0.14	0.08	0.08	0.12	0.09	0.06	0.09
K2O	7.06	8.18	7.71	9.05	8.84	9.24	8.50	8.37
Total	96.40	96.67	96.14	96.30	97.30	97.17	96.76	97.07
	Cations per 24 Oxygens							
Si	5.840	5.750	5.869	5.835	5.776	5.747	5.798	5.970
Ti	0.153	0.153	0.171	0.281	0.284	0.279	0.275	0.294
Al	3.699	3.758	3.689	3.723	3.724	3.731	3.680	3.749
Fe	2.989	2.942	2.863	3.736	3.823	3.863	3.865	3.504
Mn	0.019	0.018	0.016	0.036	0.034	0.031	0.037	0.030
Mg	2.702	2.715	2.678	1.411	1.463	1.439	1.498	1.410
Ca	0.000	0.000	0.007	0.000	0.000	0.000	0.000	0.000
Na	0.028	0.044	0.025	0.026	0.039	0.029	0.019	0.028
K	1.472	1.713	1.611	1.963	1.903	2.000	1.839	1.773
Total	16.906	17.095	16.932	17.015	17.048	17.122	17.014	16.761

APPENDIX C-3 Microprobe Analyses of Biotite (cont.)

Sample No	462Y Bt 6	462Y Bt 7	8978 Bt 1	8978 Bt 2	8978 Bt 3
Domain No	5	5	5	5	5
SiO2	34.70	34.29	34.17	34.22	34.92
TiO2	2.40	2.32	2.08	2.41	2.10
Al2O3	18.08	18.07	18.48	18.45	18.26
FeO	26.36	26.91	25.32	25.10	25.30
MnO	0.27	0.28	0.31	0.30	0.25
MgO	6.07	6.05	6.50	6.35	6.33
CaO	0.00	0.00	0.00	0.00	0.00
Na2O	0.05	0.07	0.05	0.05	0.05
K2O	8.99	8.86	9.13	9.19	9.14
Total	96.92	96.85	96.04	96.07	96.35
Cations per 24 Oxygens					
Si	5.861	5.815	5.810	5.811	5.903
Ti	0.304	0.295	0.266	0.307	0.267
Al	3.600	3.613	3.704	3.694	3.639
Fe	3.723	3.817	3.600	3.565	3.576
Mn	0.038	0.040	0.044	0.043	0.035
Mg	1.528	1.529	1.647	1.607	1.594
Ca	0.000	0.000	0.000	0.000	0.000
Na	0.016	0.023	0.016	0.016	0.016
K	1.937	1.917	1.980	1.991	1.971
Total	17.010	17.051	17.070	17.037	17.004

APPENDIX C-4 Microprobe Analyses of Plagioclase

Sample No	475A Plag 1 Core	475A Plag 1 Rim	475A Plag 2 Core	475A Plag 2 Rim	475A Plag 3 Core	475A Plag 3 Rim	475A Plag 4 Core	475A Plag 4 Rim
Domain No	2	2	2	2	2	2	2	2
SiO2	61.28	60.86	61.55	61.44	61.09	60.05	60.40	60.43
Al2O3	24.92	24.97	24.76	24.95	24.35	24.68	24.70	25.12
FeO	0.00	0.03	0.00	0.01	0.04	0.02	0.00	0.00
CaO	6.09	6.38	6.34	6.33	5.92	6.35	6.32	6.49
Na2O	8.26	8.26	8.08	8.12	8.16	7.90	8.43	8.51
K2O	0.05	0.06	0.05	0.04	0.05	0.05	0.05	0.05
Total	100.60	100.56	100.78	100.89	99.61	99.05	99.90	100.60
Cations per 32 Oxygens								
Si	10.822	10.772	10.849	10.821	10.887	10.780	10.770	10.712
Al	5.188	5.210	5.145	5.180	5.116	5.223	5.192	5.249
Fe	0.000	0.004	0.000	0.001	0.005	0.003	0.000	0.000
Ca	1.152	1.210	1.197	1.194	1.130	1.221	1.207	1.232
Na	2.828	2.834	2.761	2.772	2.819	2.749	2.914	2.925
K	0.011	0.013	0.011	0.008	0.011	0.011	0.011	0.011
Total	20.003	20.046	19.964	19.979	19.970	19.988	20.096	20.131
An	28.867	29.815	30.161	30.041	28.536	30.669	29.213	29.568
Ab	70.851	69.851	69.556	69.734	71.178	69.044	70.513	70.160
Or	0.280	0.332	0.282	0.223	0.285	0.286	0.273	0.271

APPENDIX C-4 Microprobe Analyses of Plagioclase (cont.)

Sample No	475A Plag 5 Core	475A PLag 5 Rim	476 Plag 1 Core	476 Plag 1 Rim	476 Plag 2 Core	476 Plag 2 Rim	476 Plag 3 Core	476 Plag 3 Rim
Domain No	2	2	2	2	2	2	2	2
SiO2	60.41	60.10	60.75	60.42	61.47	61.49	60.64	60.95
Al2O3	25.01	25.34	24.71	24.94	24.58	24.70	24.42	24.66
FeO	0.00	0.05	0.04	0.22	0.03	0.02	0.00	0.01
CaO	6.38	6.72	6.02	6.24	5.78	5.80	6.17	6.38
Na2O	8.51	8.34	8.16	8.10	8.61	8.44	8.33	8.09
K2O	0.06	0.06	0.05	0.44	0.06	0.04	0.05	0.04
Total	100.37	100.61	99.73	100.36	100.53	100.49	99.61	100.13
				Cations per 32 Oxygens				
Si	10.729	10.662	10.822	10.744	10.867	10.865	10.829	10.822
Al	5.236	5.299	5.189	5.228	5.122	5.145	5.141	5.162
Fe	0.000	0.007	0.005	0.032	0.004	0.002	0.000	0.001
Ca	1.214	1.277	1.149	1.189	1.094	1.098	1.180	1.213
Na	2.930	2.868	2.818	2.792	2.951	2.891	2.884	2.785
K	0.013	0.013	0.011	0.099	0.013	0.009	0.011	0.009
Total	20.124	20.129	19.997	20.087	20.053	20.012	20.047	19.993
An	29.198	30.709	28.879	29.130	26.969	27.461	28.963	30.284
Ab	70.476	68.966	70.836	68.424	72.697	72.313	70.759	69.490
Or	0.324	0.324	0.284	2.445	0.332	0.225	0.277	0.224

APPENDIX C-4 Microprobe Analyses of Plagioclase (cont.)

Sample No	476 Plag 4 Core	476 Plag 4 Rim	476 Plag 5 Core	476 Plag 5 Rim	476 Plag 6 Core	476 Plag 6 Rim
Domain No	2	2	2	2	2	2
SiO2	60.59	60.60	60.16	60.85	60.73	60.29
Al2O3	24.88	25.03	24.67	25.02	25.01	25.12
FeO	0.00	0.00	0.01	0.04	0.02	0.00
CaO	6.44	6.75	6.45	6.56	6.51	6.57
Na2O	8.25	8.09	8.34	8.14	8.30	8.15
K2O	0.11	0.04	0.05	0.04	0.06	0.06
Total	100.27	100.51	99.68	100.65	100.63	100.19
			Cations per 32 Oxygens			
Si	10.762	10.739	10.755	10.762	10.750	10.719
Al	5.210	5.229	5.199	5.217	5.219	5.265
Fe	0.000	0.000	0.001	0.005	0.002	0.000
Ca	1.225	1.281	1.235	1.243	1.234	1.251
Na	2.841	2.779	2.891	2.791	2.848	2.809
K	0.024	0.009	0.011	0.009	0.013	0.013
Total	20.065	20.040	20.095	20.029	20.070	20.059
An	29.953	31.488	29.859	30.744	30.138	30.716
Ab	69.437	68.290	69.865	69.033	69.532	68.949
Or	0.608	0.221	0.275	0.222	0.329	0.333

APPENDIX D-1 Temperature calculations for garnet-biotite pairs
Ratelfontein suite, Domain 1
(Calculation after Ferry & Spear (1978), equation 7)

Sample No 592D (Core)

Ga-Bt Pair	lnKD	Temperature °C at: 5000 bars	6000 bars	7000 bars
Ga 1 -Bt 1	-1.9958	496.1	499.6	503.0
Ga 2 -Bt 2	-2.0331	485.9	489.3	492.7
Ga 3 -Bt 3	-2.0438	483.0	486.4	489.8
Mean Ga 1-3	-2.0242	488.3	491.7	495.1

Sample No 592D (Rim)

Ga-Bt Pair	lnKD	Temperature °C at: 5000 bars	6000 bars	7000 bars
Ga 1 -Bt 1	-1.9923	497.1	500.5	504.0
Ga 2 -Bt 2	-2.0060	493.3	496.7	500.1
Ga 3 -Bt 3	-1.9624	505.5	508.9	512.4
Mean Ga 1-3	-1.9869	498.6	502.0	505.5

Sample No 593A (Core)

Ga-Bt Pair	lnKD	Temperature °C at: 5000 bars	6000 bars	7000 bars
Ga 1 -Bt 1	-1.9813	500.1	503.6	507.1
Ga 2 -Bt 2	-2.3763	403.4	406.5	409.5
Ga 3 -Bt 1	-2.1724	450.1	453.4	456.6
Ga 4 -Bt 2	-2.0786	473.8	477.2	480.5
Mean Ga 1-4	-2.1521	456.8	460.1	463.4

Sample No 593A (Rim)

Ga-Bt Pair	lnKD	Temperature °C at: 5000 bars	6000 bars	7000 bars
Ga 1 -Bt 1	-1.8343	543.6	547.2	550.9
Ga 2 -Bt 2	-2.0076	492.8	496.3	499.7
Ga 3 -Bt 1	-1.8840	528.4	532.0	535.6
Ga 4 -Bt 2	-1.8995	523.7	527.3	530.9
Mean Ga 1-4	-1.9063	522.1	525.7	529.2
Mean	Core	473	476	479
	Rim	510	514	517

APPENDIX D-2 Temperature calculations for garnet-biotite pairs
Ratelfontein suite, Domain 1
(Calculation after Indares & Martignole (1985), equation 19)

Sample No 592D (core)

Ga-Bt Pair	lnKD	Temperature °C at: 5000 bars	6000 bars	7000 bars
Ga 1 -Bt 1	-1.9958	455.5	459.0	462.4
Ga 2 -Bt 2	-2.0331	446.0	449.4	452.8
Ga 3 -Bt 3	-2.0438	448.4	451.8	455.2
Mean Ga 1-3	-2.0242	449.9	453.4	456.8

Sample No 592D (Rim)

Ga-Bt Pair	lnKD	Temperature °C at: 5000 bars	6000 bars	7000 bars
Ga 1 -Bt 1	-1.9923	448.9	452.3	455.8
Ga 2 -Bt 2	-2.0060	451.6	455.1	458.5
Ga 3 -Bt 3	-1.9624	457.9	461.4	464.9
Mean Ga 1-3	-1.9869	452.8	456.2	459.7

Sample No 593A (Core)

Ga-Bt Pair	lnKD	Temperature °C at: 5000 bars	6000 bars	7000 bars
Ga 1 -Bt 1	-1.9813	421.5	424.9	428.4
Ga 2 -Bt 2	-2.3763	372.0	375.1	378.1
Ga 3 -Bt 1	-2.1724	400.5	403.8	407.0
Ga 4 -Bt 2	-2.0786	400.8	404.1	407.5
Mean Ga 1-4	-2.1521	398.7	401.9	405.2

Sample No 593A (Rim)

Ga-Bt Pair	lnKD	Temperature °C at: 5000 bars	6000 bars	7000 bars
Ga 1 -Bt 1	-1.8343	459.1	462.8	466.4
Ga 2 -Bt 2	-2.0076	418.4	421.8	425.3
Ga 3 -Bt 1	-1.8840	458.4	462.0	465.6
Ga 4 -Bt 2	-1.8995	447.8	451.4	454.9
Mean Ga 1-4	-1.9063	445.9	449.5	453.0

Mean	Core	424	428	431
	Rim	449	453	456

APPENDIX D-3 Temperature calculations for garnet-biotite pairs
Chabiesies suite, Domain 2
(Calculation after Ferry & Spear (1978), equation 7)

Sample No 475A (Core)

Ga-Bt Pair	lnKD	Temperature °C at: 5000 bars	6000 bars	7000 bars
Ga 1 -Bt 1	-1.7049	586.1	589.9	593.8
Ga 2 -Bt 2	-1.9559	507.3	510.8	514.3
Ga 3 -Bt 3	-1.5973	625.0	629.0	633.0
Ga 4 -Bt 4	-1.5699	635.4	639.5	643.6
Ga 5 -Bt 5	-1.4173	698.4	702.8	707.1
Ga 6 -Bt 5	-1.4718	674.9	679.2	683.4
Mean Ga 1-6	-1.6195	621.1	625.2	629.2

Sample No 475A (Rim)

Ga-Bt Pair	lnKD	Temperature °C at: 5000 bars	6000 bars	7000 bars
Ga 1 -Bt 1	-1.5964	625.3	629.3	633.3
Ga 2 -Bt 2	-1.8363	543.0	546.6	550.3
Ga 3 -Bt 3	-1.7088	584.7	588.6	592.4
Ga 4 -Bt 4	-1.9035	522.6	526.1	529.7
Ga 5 -Bt 5	-1.4008	705.8	710.2	714.6
Ga 6 -Bt 5	-1.4120	700.8	705.2	709.5
Mean Ga 1-6	-1.6429	613.7	617.6	621.6

Sample No 476 (Core)

Ga-Bt Pair	lnKD	Temperature °C at: 5000 bars	6000 bars	7000 bars
Ga 1 -Bt 1	-1.4386	689.1	693.4	697.7
Ga 2 -Bt 2	-1.5998	624.0	628.0	632.0
Ga 3 -Bt 3	-1.3816	714.5	718.9	723.3
Ga 4 -Bt 1	-1.4305	692.7	697.0	701.3
Ga 5 -Bt 2	-1.4457	686.1	690.4	694.6
Ga 6 -Bt 3	-1.5042	661.5	665.7	669.9
Mean Ga 1-6	-1.4667	677.9	682.2	686.4

Sample No 476 (Rim)

Ga-Bt Pair	lnKD	Temperature °C at: 5000 bars	6000 bars	7000 bars
Ga 1 -Bt 1	-1.6243	614.9	618.9	622.8
Ga 2 -Bt 2	-1.6388	609.5	613.5	617.4
Ga 3 -Bt 3	-1.5277	652.0	656.1	660.3
Ga 4 -Bt 1	-1.5386	647.7	651.8	655.9
Ga 5 -Bt 2	-1.4789	672.0	676.2	680.4
Ga 6 -Bt 3	-1.5790	631.9	636.0	640.0
Mean Ga 1-6	-1.5645	638.0	642.0	646.1

Mean	Core	650	654	658
	Rim	626	630	634

APPENDIX D-4 Temperature calculations for garnet-biotite pairs
Chabiesies suite, Domain 2
(Calculation after Indares & Martignole (1985), equation 19)

Sample No 475A (Core)

Ga-Bt Pair	lnKD	Temperature °C at: 5000 bars	6000 bars	7000 bars
Ga 1 -Bt 1	-1.7049	513.9	517.7	521.6
Ga 2 -Bt 2	-1.9559	441.5	445.0	448.5
Ga 3 -Bt 3	-1.5973	549.3	553.3	557.3
Ga 4 -Bt 4	-1.5699	557.6	561.7	565.7
Ga 5 -Bt 5	-1.4173	619.1	623.5	627.8
Ga 6 -Bt 5	-1.4718	596.6	600.8	605.1
Mean Ga 1-6	-1.6195	546.3	550.3	554.3

Sample No 475A (Rim)

Ga-Bt Pair	lnKD	Temperature °C at: 5000 bars	6000 bars	7000 bars
Ga 1 -Bt 1	-1.5964	547.8	551.8	555.8
Ga 2 -Bt 2	-1.8363	472.7	476.4	480.0
Ga 3 -Bt 3	-1.7088	513.3	517.1	521.0
Ga 4 -Bt 4	-1.9035	458.8	462.4	465.9
Ga 5 -Bt 5	-1.4008	623.7	628.1	632.5
Ga 6 -Bt 5	-1.4120	618.5	622.8	627.2
Mean Ga 1-6	-1.6429	539.1	543.1	547.0

Sample No 476 (Core)

Ga-Bt Pair	lnKD	Temperature °C at: 5000 bars	6000 bars	7000 bars
Ga 1 -Bt 1	-1.4386	609.4	613.7	618.0
Ga 2 -Bt 2	-1.5998	550.9	554.9	558.9
Ga 3 -Bt 3	-1.3816	623.1	627.5	631.9
Ga 4 -Bt 1	-1.4305	610.0	614.3	618.6
Ga 5 -Bt 2	-1.4457	605.6	609.9	614.2
Ga 6 -Bt 3	-1.5042	576.5	580.7	584.9
Mean Ga 1-6	-1.4667	595.9	600.1	604.4

Sample No 476 (Rim)

Ga-Bt Pair	lnKD	Temperature °C at: 5000 bars	6000 bars	7000 bars
Ga 1 -Bt 1	-1.6243	541.9	545.8	549.8
Ga 2 -Bt 2	-1.6388	537.8	541.7	545.7
Ga 3 -Bt 3	-1.5277	572.0	576.1	580.2
Ga 4 -Bt 1	-1.5386	571.4	575.6	579.7
Ga 5 -Bt 2	-1.4789	594.8	599.0	603.2
Ga 6 -Bt 3	-1.5790	551.4	555.4	559.5
Mean Ga 1-6	-1.5645	561.5	565.6	569.7

Mean	Core	571	575	579
	Rim	550	554	558

APPENDIX D-5 Temperature calculations for garnet-biotite pairs
Ratelfontein suite, Domain 4
(Calculation after Ferry & Spear (1978), equation 7)

Sample No 317A (Core)

Ga-Bt Pair	lnKD	Temperature °C at: 3000 bars	3500 bars	4000 bars
Ga 1 -Bt 1	-1.4677	668.2	670.3	672.4
Ga 2 -Bt 2	-1.3847	704.2	706.5	708.7
Ga 3 -Bt 3	-1.5115	650.2	652.3	654.4
Ga 4 -Bt 4	-1.4279	685.2	687.3	689.5
Ga 5 -Bt 5	-1.4892	659.3	661.4	663.5
Ga 6 -Bt 6	-1.3062	741.0	743.3	745.6
Mean Ga 1-6	-1.4312	684.6	686.8	689.0

Sample No 317A (Rim)

Ga-Bt Pair	lnKD	Temperature °C at: 3000 bars	3500 bars	4000 bars
Ga 1 -Bt 1	-1.2625	762.7	765.0	767.3
Ga 2 -Bt 2	-1.2771	755.3	757.7	760.0
Ga 3 -Bt 3	-1.2655	761.2	763.5	765.8
Ga 4 -Bt 4	-1.2716	758.1	760.4	762.7
Ga 5 -Bt 5	-1.3121	738.1	740.4	742.7
Ga 6 -Bt 6	-1.2993	744.4	746.6	748.9
Mean Ga 1-6	-1.2813	753.3	755.6	757.9

Mean	Core	685	687	689
	Rim	753	756	758

APPENDIX D-6 Temperature calculations for garnet-biotite pairs
Ratelfontein suite, Domain 4
(Calculation after Indares & Martignole (1985), equation 19)

Sample No 317A (CORE)

Ga-Bt Pair	lnKD	Temperature °C at: 3000 bars	3500 bars	4000 bars
Ga 1 -Bt 1	-1.4677	585.4	587.5	589.6
Ga 2 -Bt 2	-1.3847	589.9	592.1	594.3
Ga 3 -Bt 3	-1.5115	567.7	569.7	571.8
Ga 4 -Bt 4	-1.4279	589.3	591.4	593.6
Ga 5 -Bt 5	-1.4892	577.3	579.4	581.5
Ga 6 -Bt 6	-1.3062	628.6	630.9	633.2
Mean Ga 1-6	-1.4312	589.7	591.8	594.0

Sample No 317A (RIM)

Ga-Bt Pair	lnKD	Temperature °C at: 3000 bars	3500 bars	4000 bars
Ga 1 -Bt 1	-1.2625	667.0	669.3	671.7
Ga 2 -Bt 2	-1.2771	634.3	636.7	639.0
Ga 3 -Bt 3	-1.2655	647.1	649.5	651.8
Ga 4 -Bt 4	-1.2716	653.4	655.7	658.0
Ga 5 -Bt 5	-1.3121	637.2	639.5	641.7
Ga 6 -Bt 6	-1.2993	631.1	633.3	635.6
Mean Ga 1-6	-1.2814	645.0	647.3	649.6

Mean	Core	590	592	594
	Rim	645	647	650

APPENDIX D-7 Temperature calculations for garnet-biotite pairs
Chabiesies suite, Domain 5
(Calculation after Ferry & Spear (1978), equation 7)

Sample No 462Y (Core)

Ga-Bt Pair	lnKD	Temperature °C at: 3000 bars	3500 bars	4000 bars
Ga 1 -Bt 1	-1.0237	899.6	902.3	904.9
Ga 2 -Bt 2	-1.1987	796.0	798.4	800.9
Ga 3 -Bt 3	-1.2175	786.0	788.4	790.7
Ga 4 -Bt 4	-0.9760	931.4	934.2	936.9
Ga 5 -Bt 5	-1.3071	740.5	742.8	745.1
Ga 6 -Bt 6	-1.0139	906.0	908.7	911.4
Mean Ga 1-6	-1.1228	843.0	845.8	848.3

Sample No 462Y (Rim)

Ga-Bt Pair	lnKD	Temperature °C at: 3000 bars	3500 bars	4000 bars
Ga 1 -Bt 1	-1.0962	854.4	856.9	859.5
Ga 2 -Bt 2	-1.3643	713.5	715.8	718.0
Ga 3 -Bt 3	-1.3012	743.4	745.7	748.0
Ga 4 -Bt 4	-1.1182	841.3	843.8	846.3
Ga 5 -Bt 5	-1.3040	742.1	744.4	746.6
Ga 6 -Bt 6	-1.1342	832.0	834.5	837.0
Mean Ga 1-6	-1.2196	787.7	790.1	792.6

Sample No 8978 (Core)

Ga-Bt Pair	lnKD	Temperature °C at: 3000 bars	3500 bars	4000 bars
Ga 1 -Bt 1	-1.1352	831.4	833.9	836.4
Ga 2 -Bt 2	-1.1586	818.1	820.6	823.1
Ga 3 -Bt 3	-1.3036	742.2	744.5	746.8
Mean Ga 1-3	-1.1991	797.2	799.6	802.1

Sample No 8978 (Rim)

Ga-Bt Pair	lnKD	Temperature °C at: 3000 bars	3500 bars	4000 bars
Ga 1 -Bt 1	-1.2390	774.7	777.1	779.4
Ga 2 -Bt 2	-1.1910	800.2	802.6	805.1
Ga 3 -Bt 3	-1.3638	713.8	716.0	718.2
Mean Ga 1-3	-1.2646	762.9	765.2	767.5

Mean	Core	820	823	825
	Rim	775	778	780

APPENDIX D-8 Temperature calculations for garnet-biotite pairs
Chabiesies suite, Domain 5
(Calculation after Indares & Martignole (1985), equation 19)

Sample No 462Y (Core)

Ga-Bt Pair	lnKD	Temperature °C at: 3000 bars	3500 bars	4000 bars
Ga 1 -Bt 1	-1.0237	823.1	825.7	828.3
Ga 2 -Bt 2	-1.1987	763.8	766.2	768.6
Ga 3 -Bt 3	-1.2175	760.6	763.0	765.4
Ga 4 -Bt 4	-0.976	847.8	850.5	853.2
Ga 5 -Bt 5	-1.3071	687.8	690.1	692.4
Ga 6 -Bt 6	-1.0139	811.8	814.4	817.1
Mean Ga 1-6	-1.1228	782.5	784.9	787.5

Sample No 462Y (RIM)

Ga-Bt Pair	lnKD	Temperature °C at: 3000 bars	3500 bars	4000 bars
Ga 1 -Bt 1	-1.0962	798.4	801.0	803.5
Ga 2 -Bt 2	-1.3643	701.7	703.9	706.2
Ga 3 -Bt 3	-1.3012	724.7	727.0	729.3
Ga 4 -Bt 4	-1.1182	782.1	784.7	787.2
Ga 5 -Bt 5	-1.304	689.7	692.0	694.3
Ga 6 -Bt 6	-1.1342	759.9	762.4	764.9
Mean Ga 1-6	-1.2197	742.7	745.1	747.5

Sample No 8978 (Core)

Ga-Bt Pair	lnKD	Temperature °C at: 3000 bars	3500 bars	4000 bars
Ga 1 -Bt 1	-1.1352	803.1	805.6	808.1
Ga 2 -Bt 2	-1.1586	775.0	777.5	779.9
Ga 3 -Bt 3	-1.3036	755.5	757.8	760.1
Mean Ga 1-3	-1.1991	777.8	780.3	782.7

Sample No 8978 (Rim)

Ga-Bt Pair	lnKD	Temperature °C at: 3000 bars	3500 bars	4000 bars
Ga 1 -Bt 1	-1.2390	797.2	799.6	801.9
Ga 2 -Bt 2	-1.1910	766.2	768.6	771.1
Ga 3 -Bt 3	-1.3638	731.6	733.8	736.1
Mean Ga 1-3	-1.2646	765.0	767.3	769.7

Mean	Core	780	783	785
	Rim	754	756	759

APPENDIX D-9 Pressure calculations for garnet-plagioclase pairs Chabiesies suite, Domain 2 – Pan African event (Calculation after Waters; in Kasch (1983), equation 27)

Sample No 475A (Core)

Ga-Plag Pair	lnK	Pressure (bars) at: 500 °C	600 °C	700 °C
Ga 1 -Pl 1	-6.2697	3366.8	5048.9	6731.0
Ga 2 -Pl 2	-6.4187	3222.0	4885.4	6548.8
Ga 3 -Pl 3	-6.2017	3432.7	5123.4	6814.0
Ga 4 -Pl 4	-5.7824	3840.0	5583.3	7326.6
Ga 5 -Pl 5	-6.2538	3382.2	5066.3	6750.4
Ga 6 -Pl 5	-6.2127	3422.1	5111.3	6800.6
Mean Ga 1-6	-6.1898	3444.3	5136.4	6828.5

Sample No 475A (Rim)

Ga-Plag Pair	lnK	Pressure (bars) at: 500 °C	600 °C	700 °C
Ga 1 -Pl 1	-6.3501	3288.7	4960.7	6632.7
Ga 2 -Pl 2	-6.3070	3330.5	5007.9	6685.3
Ga 3 -Pl 3	-6.5278	3116.1	4765.8	6415.5
Ga 4 -Pl 4	-6.2100	3424.7	5114.3	6803.9
Ga 5 -Pl 5	-6.5403	3103.9	4752.0	6400.2
Ga 6 -Pl 5	-6.4551	3186.7	4845.5	6504.3
Mean Ga 1-6	-6.3983	3241.7	4907.7	6573.7

Sample No 476 (Core)

Ga-Plag Pair	lnK	Pressure (bars) at: 500 °C	600 °C	700 °C
Ga 1 -Pl 1	-6.3830	3256.7	4924.5	6592.4
Ga 2 -Pl 2	-6.0992	3532.3	5235.9	6939.4
Ga 3 -Pl 3	-6.3395	3299.0	4972.3	6645.7
Ga 4 -Pl 4	-6.5628	3082.1	4727.4	6372.7
Ga 5 -Pl 5	-6.4217	3219.1	4882.1	6545.1
Ga 6 -Pl 6	-6.3654	3273.8	4943.8	6613.9
Mean Ga 1-6	-6.3619	3277.1	4947.6	6618.2

Sample No 476 (Rim)

Ga-Plag Pair	lnK	Pressure (bars) at: 500 °C	600 °C	700 °C
Ga 1 -Pl 1	-6.1341	3498.4	5197.6	6896.7
Ga 2 -Pl 2	-6.0705	3560.2	5267.3	6974.5
Ga 3 -Pl 3	-6.3720	3267.3	4936.6	6605.9
Ga 4 -Pl 4	-6.4721	3170.2	4826.9	6483.5
Ga 5 -Pl 5	-6.4173	3223.4	4887.0	6550.6
Ga 6 -Pl 6	-6.3005	3336.8	5015.1	6693.3
Mean Ga 1-6	-6.2944	3342.7	5021.7	6700.7

Mean	Core	3361	5042	6723
	Rim	3292	4965	6637

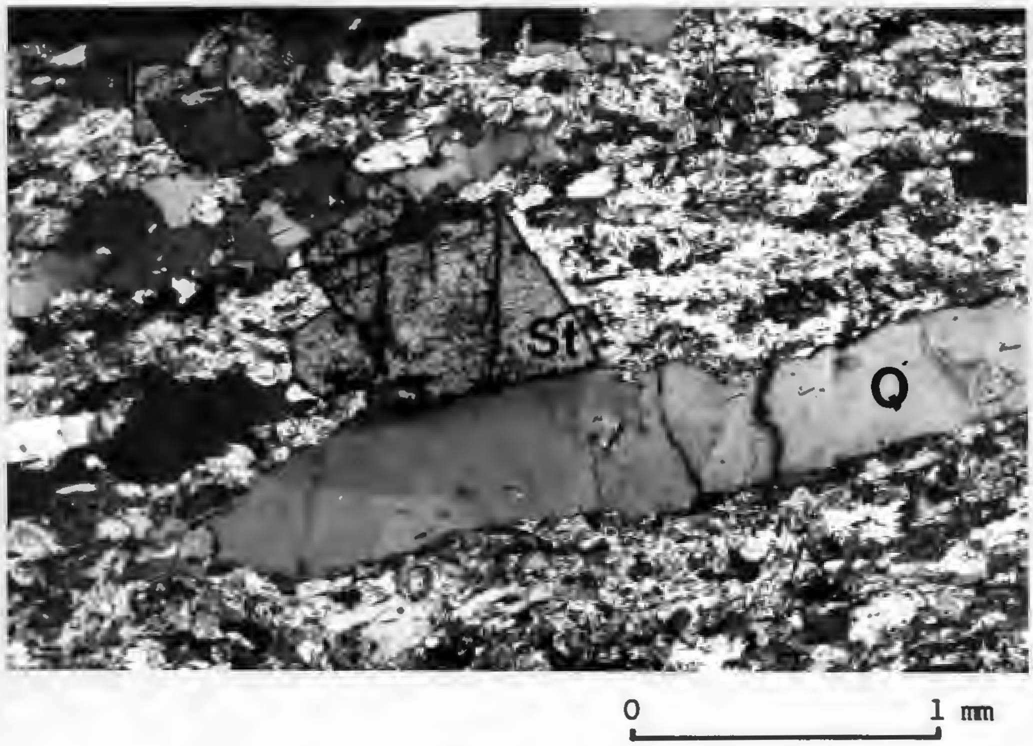


Plate 1 Photomicrograph showing texture in metapelite, Chabiesies suite. Note quartz ribbon (Q), and euhedral staurolite grain (St). The latter mineral retains S1, which is parallel to the regional schistosity.

Plate 2 Metapelite bed in banded quartzite, Ratelfontein suite. Irregular thickness and shape of bed is due to thrusting.

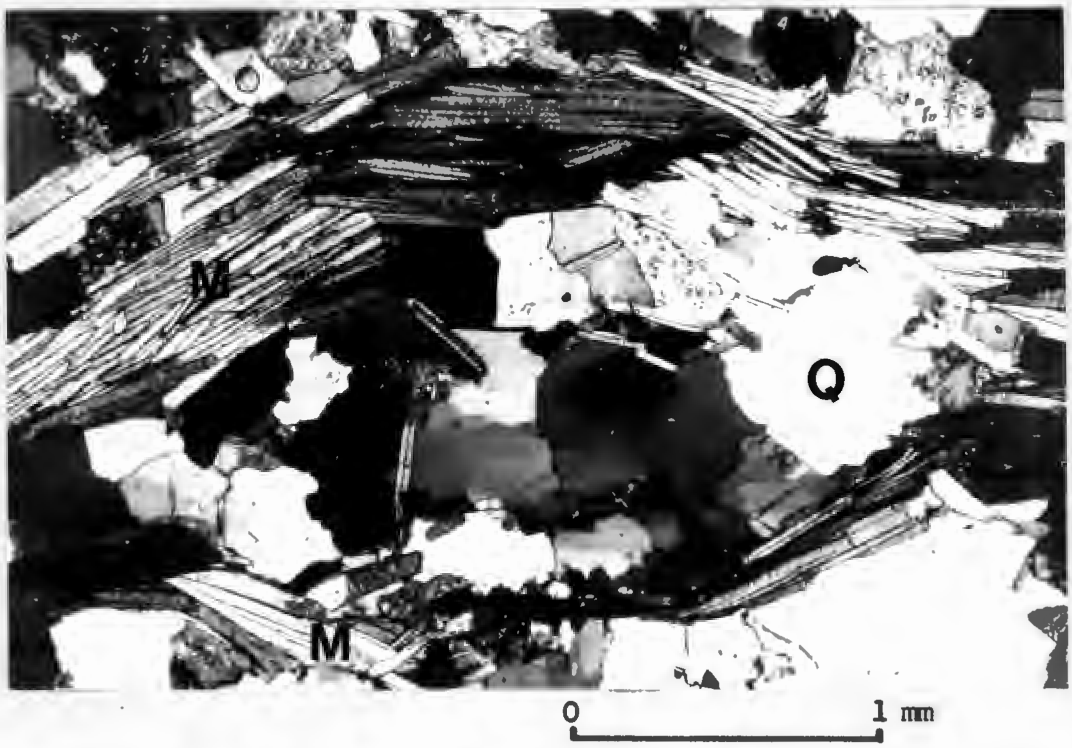


Plate 3 Photomicrograph showing sigmoidal cleavage outlined by muscovite, in a quartz-muscovite schist, Groenrivier suite. M = muscovite, Q = quartz.

Plate 4 Cigar-shaped cordierite porphyroblasts in cordierite schist forming prominent linear structures, Groenrivier suite.

Plate 5 Hornblende gneiss, displaying garbenschiefer texture, Groenrivier suite.

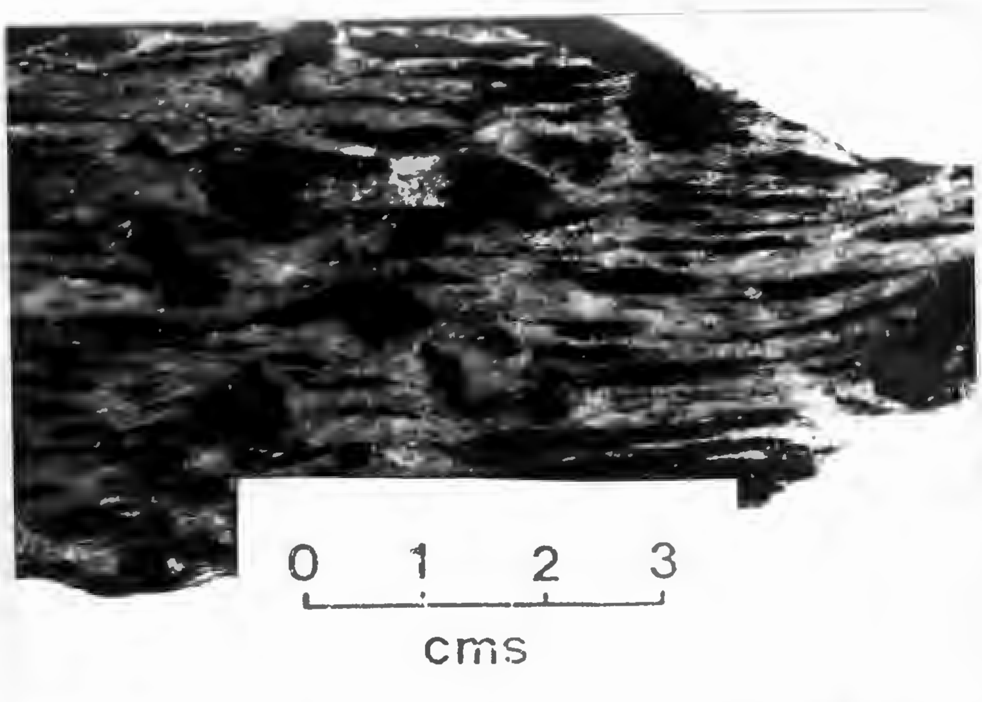


Plate 6 Hornblende gneiss at southern contact of Vioolsdrif granodiorite, Groothoek Thrust zone. Note large, randomly oriented hornblende porphyroblasts. A dextral sense of shear is interpreted from orientation of porphyroblasts.

Plate 7   Transposed layering in metavolcanics of the Windvlakte suite.

Plate 8   Mafic and ultramafic rocks of the Klipbok complex in the centre foreground, Groothoek Thrust zone, facing north-west. Nama unconformity in the background, dips towards the north.

Plate 9   The Riethoek Shear, facing south. Note dip of 65° towards the right (west).

Plate 10   The Steenbok Shear, facing north. Note near vertical foliation over a broad zone.

Plate 11 Isoclinally folded Nama quartzite within the Steenbok Shear.

Plate 12 Similar folds in Sabieboomrante adamellite gneiss adjacent to Steenbok Shear. Note slip has taken place parallel to the axial planes of similar folds.

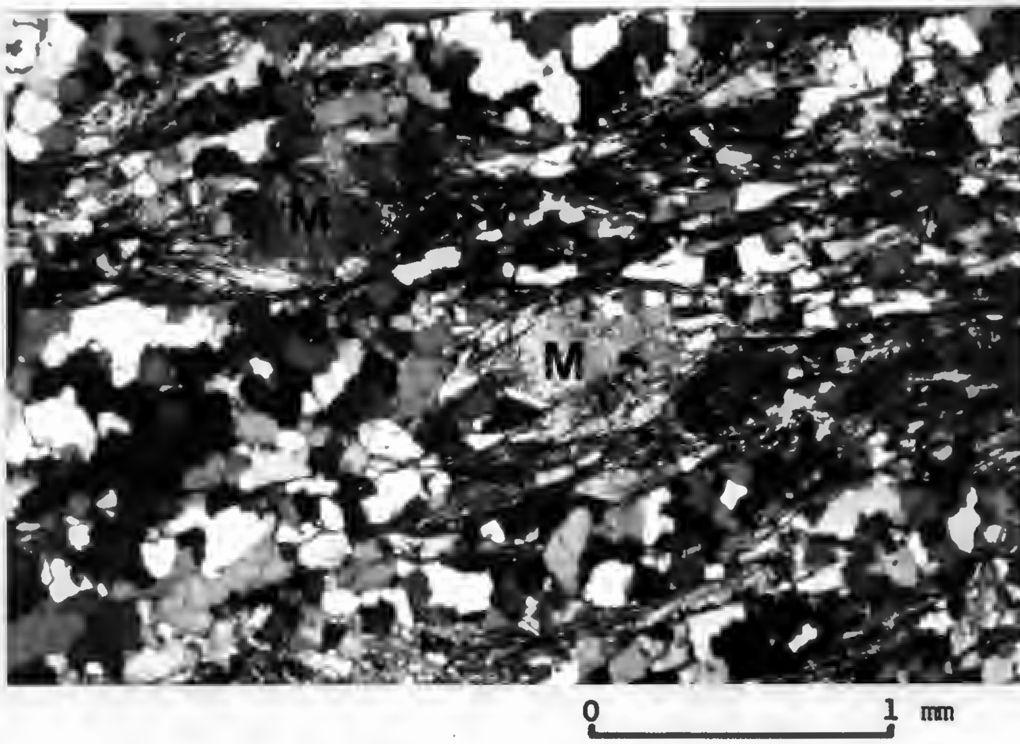


Plate 13 Photomicrograph of quartz muscovite schist showing deformation textures in the Steenbok Shear. A dextral sense of shear is interpreted from shape and orientation of muscovite "fishes" (M).

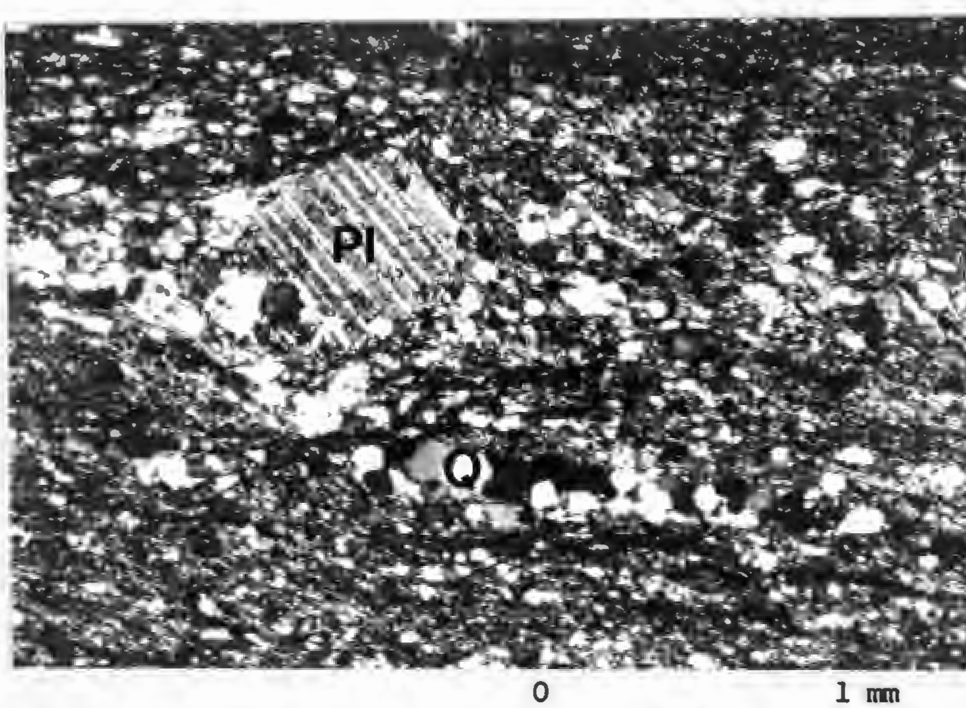


Plate 14 Photomicrograph showing mylonitic texture in the Windvlakte Thrust. Note quartz ribbons (Q) and plagioclase porphyroclast (Pl). A dextral sense of shear is indicated by orientation of porphyroclasts.

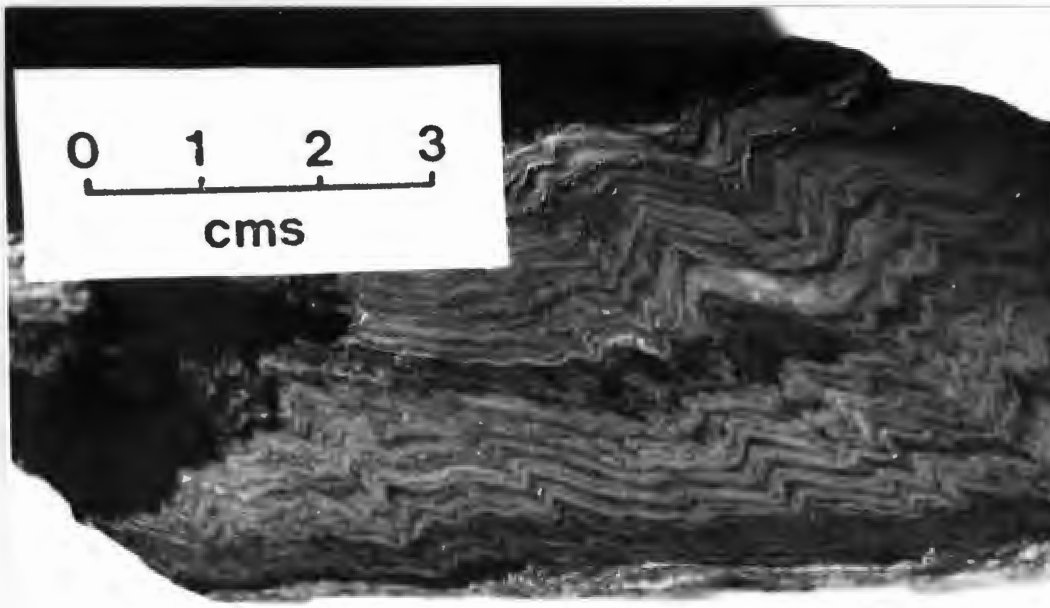


Plate 15 Chevron folds in mylonite, Windvlakte Thrust. Folds verge towards the north (left).

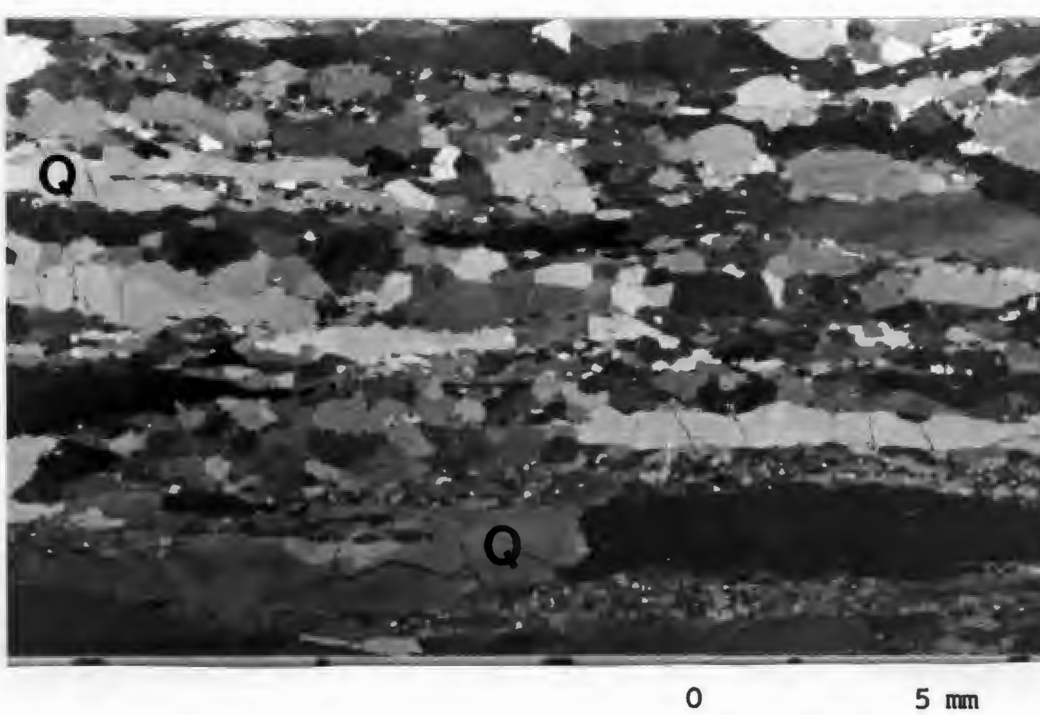


Plate 16 Mylonite texture in Leucogranite, Groothoek Thrust zone. Note prominent development of quartz ribbons (Q).

Plate 17 Thrust plane in micaceous schists of the Ratelfontein suite. Sense of movement along thrust indicated by arrows.

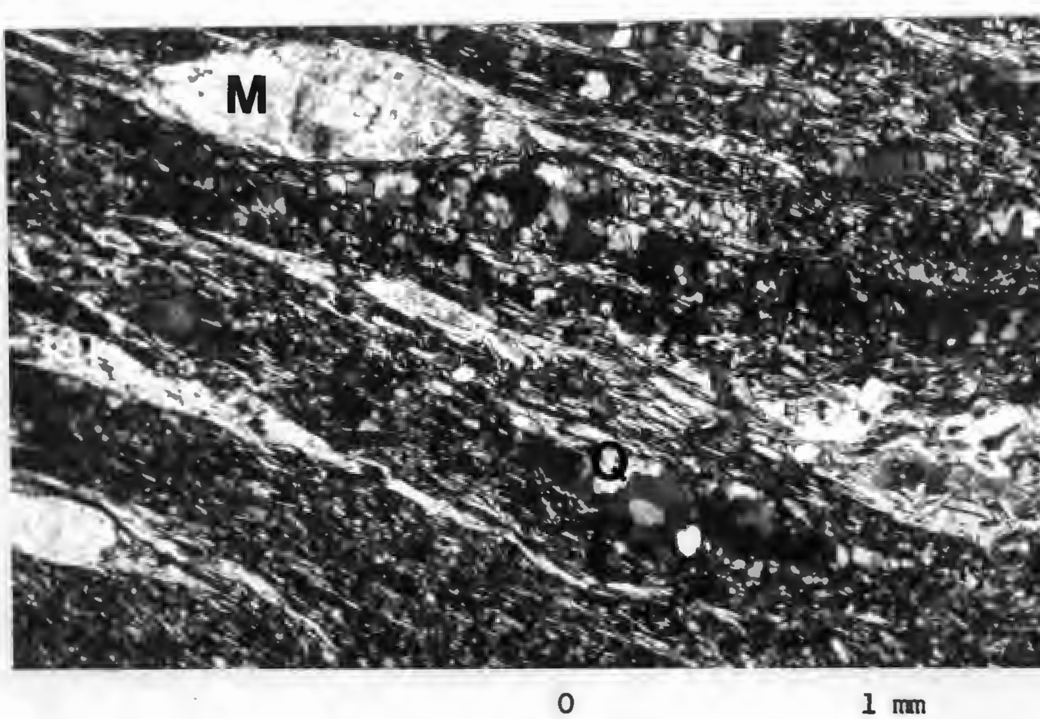


Plate 18 Photomicrograph showing "fish" micas (M) and quartz ribbons (Q) in Kouefontein North Ramp structure. A right-lateral sense of shear is interpreted from shape and orientation of micas.

Plate 19 Well developed foliation and lineation in adamellite, Sabieboomrante Thrust zone.

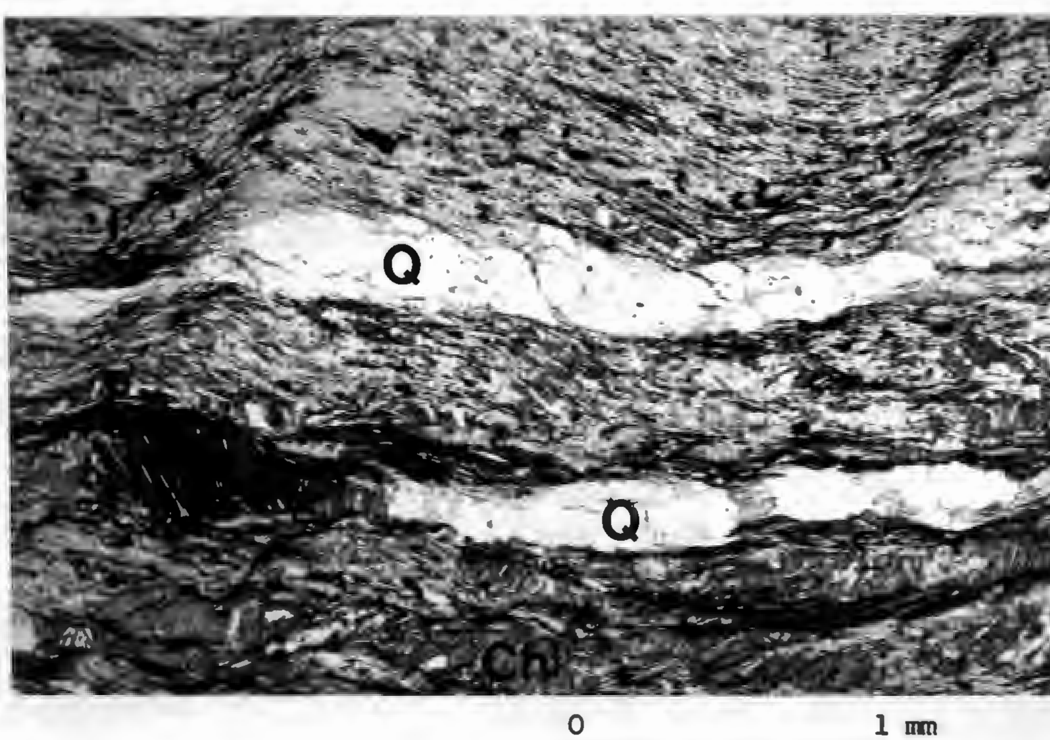


Plate 20 Photomicrograph showing mylonite texture and cleavage development in chlorite-sericite schist, Chabiesies South Thrust. Note:

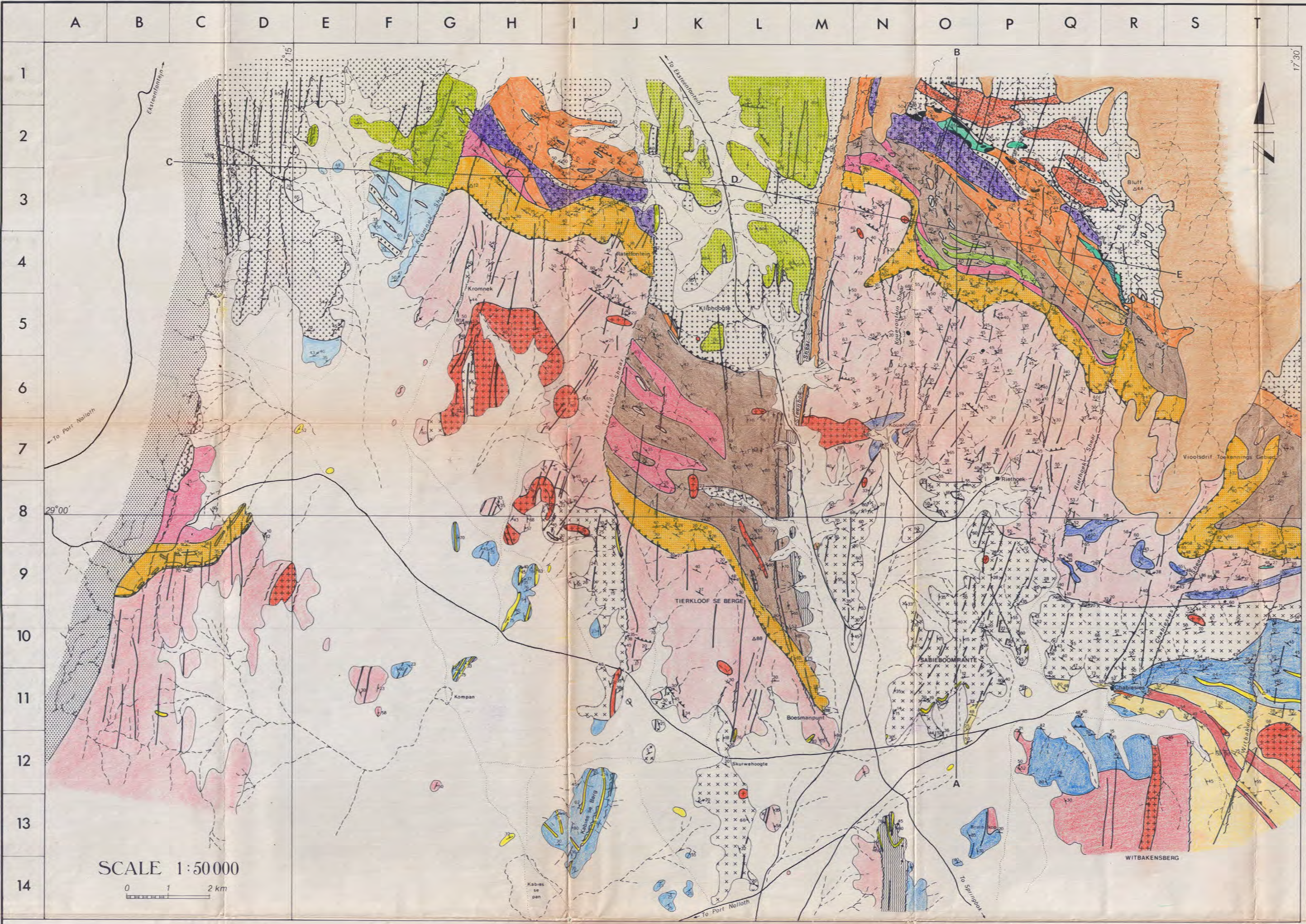
(a) boudinaged quartz ribbons (Q),

(b) skeletal intergrowths of phyllosilicates and quartz (Chl) modified into fish shapes within microlithons,

(c) crenulation cleavage in e.g. top left corner.

ANNEXURE 1

GEOLOGY OF AN AREA SOUTH - EAST OF EKSTEENFONTEIN



LOCALITY MAP

LEGEND

SEDIMENTARY AND METAVOLCANIC ROCKS

ROCK TYPES	FORMATION/ SUITE	GROUP	SUBPROVINCE
Wind-blown sand, scree			
Quartzites, shales, limestone		Nama (Kuibis Subgroup)	
Quartzites, schists	Stinkfontein	Gariep	
Quartzo-feldspathic gneisses, banded-iron formation, porphyritic gneisses	Windvlakte (upper unit)	Orange River	Richtersveld
Quartzo-feldspathic gneisses; quartz rich gneisses, biotite quartz feldspar gneisses	Windvlakte (lower unit)	Orange River	Richtersveld
Leucocratic & mesocratic quartzo-feldspathic gneisses, hornblende gneisses, biotite schists, biotite quartz feldspar gneisses	Groenrivier	Transition Zone	Richtersveld
Cordierite schist phlogopite-corundum schist	Groenrivier	Transition Zone	Richtersveld
Muscovite quartz feldspar schist	Groenrivier	Transition Zone	Richtersveld
Biotite quartz feldspar schist	Groothoek	Transition Zone	Richtersveld
Chlorite schist, muscovite quartz schist	Groothoek	Transition Zone	Richtersveld
Quartzite, magnetite quartzite, staurolite garnet schist, cordierite anthophyllite sillimanite schist, muscovite schist, biotite schist	Ratelfontein	Transition Zone	Richtersveld
Xenolithic rocks comprising muscovite schists chlorite biotite schist (± chloritoid) biotite quartz feldspar gneiss, amphibolite	Chabiesies	Een Riet Subgroup	Bushmanland
Quartzite	Chabiesies	Een Riet Subgroup	Bushmanland
Biotite gneiss and schist (± garnet chloritoid)	Chabiesies	Een Riet Subgroup	Bushmanland
Muscovite biotite quartz feldspar gneiss and schist + staurolite ± kyanite	Chabiesies	Een Riet Subgroup	Bushmanland

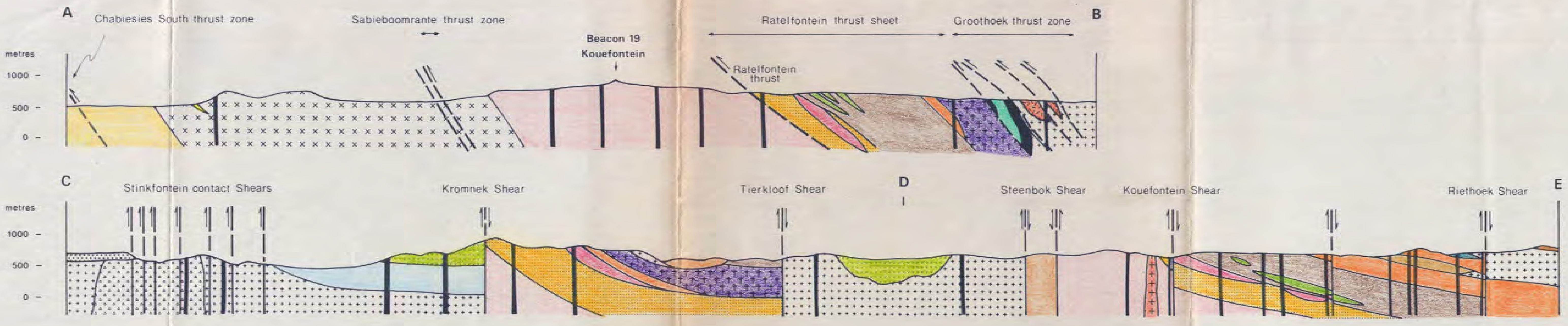
TECTONIZED ROCK

- Phyllonite

INTRUSIVE ROCKS

Mafic dyke, Gannakouriep Dyke Suite	
Large pegmatite	
K-granites (post tectonic; Spektakel Suite ?)	
Even-grained quartz feldspar gneiss (Eyams granite ?)	
Leucogranite (pretectonic)	
Dabbieputs muscovite biotite granite gneiss	
Nodular quartz feldspar gneiss (+ sillimanite) (Noenoemaasberg gneiss ?)	
Finely laminated biotite gneiss (Steinkopf gneiss ?)	
Kouefontein granite gneiss	
Sabieboomrante adamellite gneiss	
Tweeriviere granite gneiss	Vioolsdrif Suite
Megacrystic granite	Vioolsdrif Suite
Vioolsdrif granodiorite	Vioolsdrif Suite
Amphibolite	Vioolsdrif Suite
Serpentinite	Vioolsdrif Suite

CROSS-SECTIONS A-B AND C-D-E



GEOLOGICAL KEY

- Limit of outcrop
- Lithological contact
- Strike and dip of foliation
- Vertical foliation
- Trend and plunge of lineation
- Zone of thrusting
- Zone of intense shearing

GEOGRAPHICAL KEY

- Water course
- Pan
- Trigonometrical beacon and number
- Secondary road
- Track
- Abandoned farmhouse

PRU

Mapped and compiled by P.W.K. Booth, 1984-1989. Drawn by S. Sayers, 1989. Precambrian Research Unit, Department of Geology, University of Cape Town. ©

GEOLOGY OF THE KLIPBOK MAFIC - ULTRAMAFIC AND ASSOCIATED ROCKS

MAP SYMBOLS

- *Water course*
- *Track*
- *Lithological contact*
- *Strike and dip of foliation* (70)
- *Vertical foliation*
- *Trend and plunge of lineation* (30)
- *Localized zones of shearing*
- *Thrust zone*
- *Major shear zone*

N

0 — 1 kms

SCALE 1: 30,000

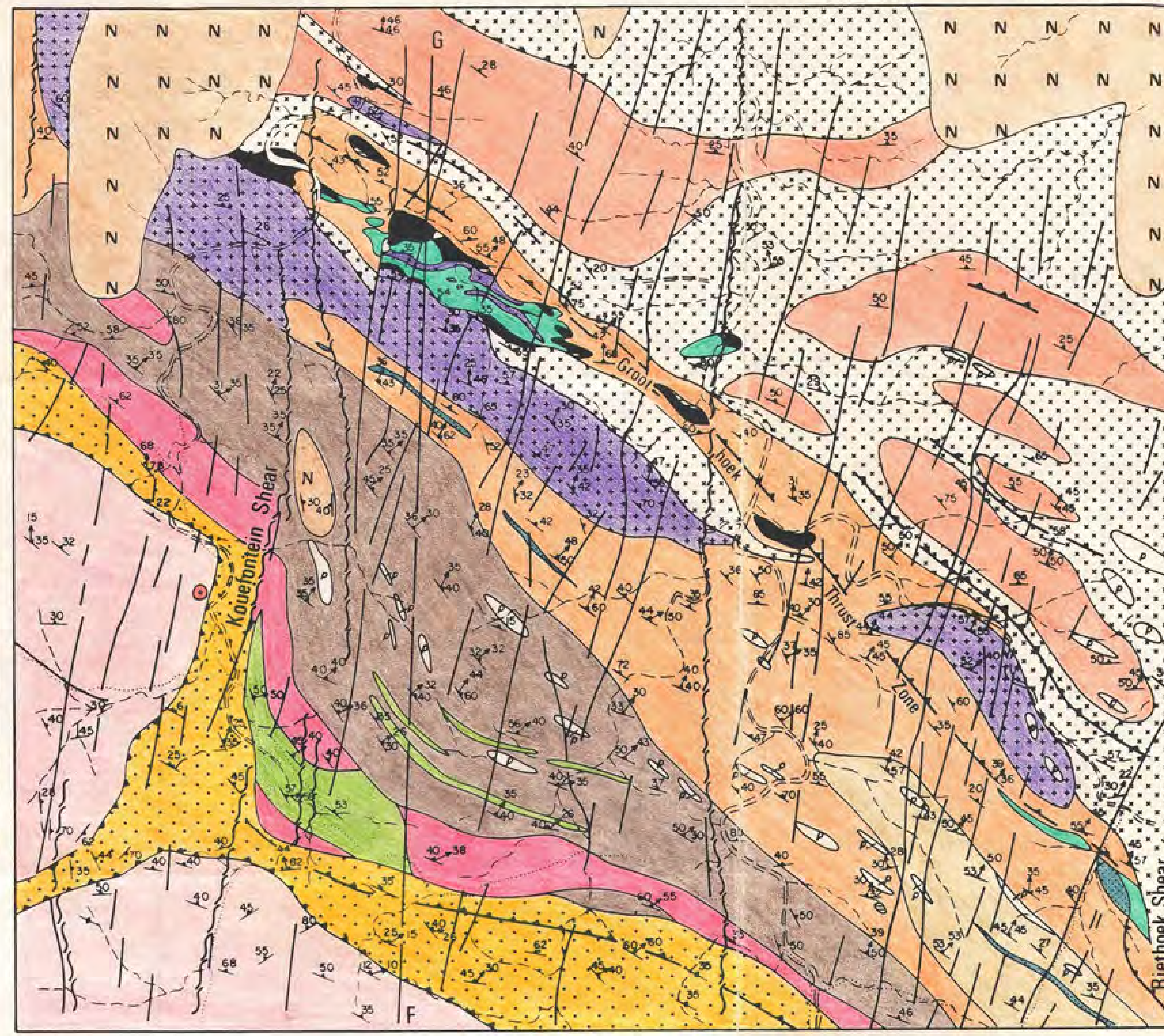
METAMORPHOSED SUPRACRUSTALS, AND COVER ROCKS

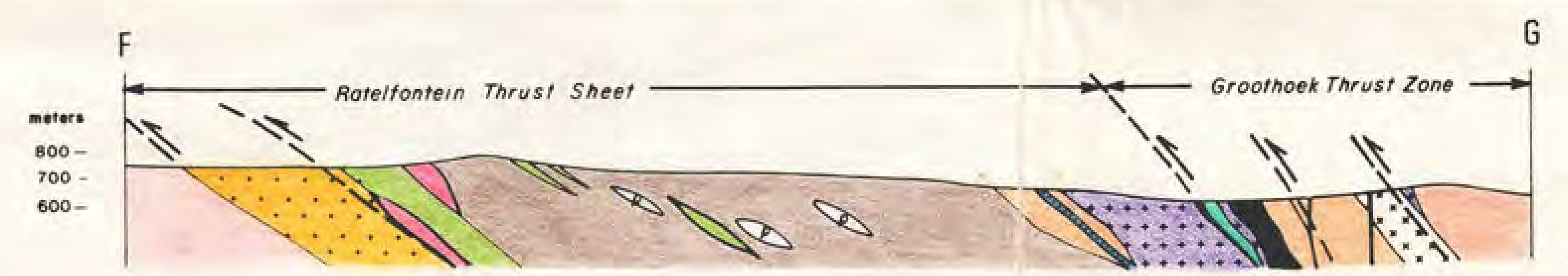
	SUITE	ROCK TYPES	
Nama Group (Kuibis Subgroup)		Quartzites, shales, limestone.	N
Richtersveld Subprovince	Windvlakte (lower unit)	Quartzo-feldspathic gneisses, quartz-rich gneisses, biotite quartz feldspar gneisses.	
	Groenrivier	Leucocratic & mesocratic quartzo-feldspathic gneisses, hornblende gneisses, biotite schists, biotite quartz feldspar gneisses.	
		Cordierite schist-phlogopite corundum schist.	
		Muscovite quartz feldspar schist.	
	Groothoek	Biotite quartz feldspar schist.	
		Chlorite schist, muscovite quartz schist, garnet biotite schist.	
	Ratelfontein	Quartzite, magnetite quartzite, staurolite garnet cordierite anthophyllite sillimanite schist, muscovite schist, biotite schist.	

INTRUSIVE ROCKS

- *Mafic dyke, Ganakouriep Suite.*
- *Large pegmatite.*
- *K-granites (syn-post tectonic) Spektakel Suite?*
- *Kouefontein granite gneiss.*
- *Leucogranite (pretectonic)*
- *Dabbieputs muscovite biotite granite gneiss.*
- *Granodiorite, Vioolsdrif Suite.*
- *Amphibolite.*
- *Hornblendite.*
- *Serpentinite.*

(Amphibolite, Hornblendite, Serpentinite: Klipbok complex)





CROSS - SECTION F-G


ANNEXURE 3
17°15'
17°30'
29°00'
DOMAIN 1
SUBDOMAIN 1.1
[a]
N
$N_{S(N)_2}$ = 58
$N_{L(N)_2}$ = 40
Contours = 4-6-8%, max 10%
[b]
$N_{S(P)_2}$ = 30
$N_{L(P)_2}$ = 10
Contours = 16-20-24%, max 30%
GARIEP
GROUP
DOMAIN 2
SUBDOMAIN 2.1
$N_{S(N)_2}$ = 67
$N_{L(N)_2}$ = 29
Contours = 4-6-8-12%, max 19%
SUBDOMAIN 2.2
$N_{S(N)_2}$ = 56
$N_{L(N)_2}$ = 10
Contours = 4-6-8-10%, max 13%
SUBDOMAIN 2.3
$N_{S(N)_2}$ = 36
$N_{L(N)_2}$ = 7
Contours = 4-8-12-16%, max 19%
SUBDOMAIN 2.4
$N_{S(N)_2}$ = 57
$N_{L(N)_2}$ = 14
Contours = 4-6-8%, max 9%
KROMNEK SHEAR
TIERKLOOF SHEAR
DOMAIN 3
SUBDOMAIN 3.1
$N_{S(P)_2}$ = 52
$N_{L(P)_2}$ = 16
Contours = 5-10-15-20%, max 23%
SUBDOMAIN 3.2
$N_{S(N)_2}$ = 76
$N_{L(N)_2}$ = 20
Contours = 6-8-10-12%, max 18%
SUBDOMAIN 3.3
$N_{S(N)_2}$ = 50
$N_{L(N)_2}$ = 20
Contours = 6-10-14%, max 20%
DOMAIN 4
SUBDOMAIN 4.1
$N_{S(P)_2}$ = 25
$N_{L(P)_2}$ = 10
Contours = 10-20-30%, max 36%
SUBDOMAIN 4.2
$N_{S(P)_2}$ = 50
$N_{L(P)_2}$ = 15
Contours = 10-15-20%, max 28%
STEENBOK SHEAR
DOMAIN 5
SUBDOMAIN 5.1
$N_{S(N)_2}$ = 32
$N_{L(N)_2}$ = 8
SUBDOMAIN 5.2
$N_{S(N)_2}$ = 68
$N_{L(N)_2}$ = 13
SUBDOMAIN 5.3
$N_{S(N)_2}$ = 39
$N_{L(N)_2}$ = 6
KOUEFONTEIN SHEAR
DOMAIN 6
SUBDOMAIN 6.1
$N_{S(N)_2}$ = 153
$N_{L(N)_2}$ = 88
Contours = 4-8-12-16%, max 22%
$N_{S(N)_2}$ = 78
$N_{L(N)_2}$ = 37
Contours = 10-12-14%, max 17%
SUBDOMAIN 6.2
SUBDOMAIN 6.3
$N_{S(N)_2}$ = 120
$N_{L(N)_2}$ = 27
Contours = 2-4-6-8%, max 12%
SUBDOMAIN 6.4
$N_{S(N)_2}$ = 58
$N_{L(N)_2}$ = 9
Contours = 4-6-8%, max 9%
NAMA GROUP
DOMAIN 7
SUBDOMAIN 7.1
$N_{S(N)_2}$ = 22
$N_{L(N)_2}$ = 7
Contours = 8-12-16-20%, max 27%
SUBDOMAIN 7.2
$N_{S(N)_2}$ = 53
$N_{L(N)_2}$ = 25
Contours = 6-8-10-12%, max 15%
RIETHOEK SHEAR
CHABIESIES SHEAR
DOMAIN 8
SUBDOMAIN 8.1
$N_{S(N)_2}$ = 61
$N_{L(N)_2}$ = 33
Contours = 2-4-6-8%, max 10%
SUBDOMAIN 8.2
$N_{S(N)_2}$ = 37
$N_{L(N)_2}$ = 10
Contours = 6-10-14-18%, max 22%
SUBDOMAIN 8.3
$N_{S(N)_2}$ = 31
$N_{L(N)_2}$ = 13
Contours = 6-10-14%, max 16%
WITBAKENSBERG SHEAR
Subdomain 5.1
Contours = 5-10-20%, max 34%
Subdomain 5.2
Contours = 2-4-6%, max 10%
Subdomain 5.3
Contours = 4-8%, max 10%
STRUCTURAL ELEMENTS OF AN AREA
SOUTH OF EKSTEENFONTEIN
Major shear zone
Subdomain boundary
Contours = density of poles to foliations
$S(N)_2$; $L(N)_2$ = Namaqua fabrics
$S(P)_2$; $L(P)_2$ = Pan-African fabrics
△ = π pole to S-surfaces
• = stretching lineation
x = tight folds
o = open/gentle folds
0 1 2 3 4 km
Scale 1:50,000